

13
51
63
0.30
Science

UNION GÉODÉSIQUE ET GÉOPHYSIQUE INTERNATIONALE

Association Internationale
d'Hydrologie Scientifique

ASSEMBLÉE GÉNÉRALE D'OSLO 19-28 août 1948

TOME II

Procès-verbaux des séances

Travaux de la Commission de la Neige et des Glaciers

U. of IL LIBRARY
JUL 7. 1971
CHICAGO CIRCLE

Imprimerie Ceuterick
66, rue Vital Decoster
LOUVAIN (Belgique)



Digitized by the Internet Archive
in 2024

UNION GÉODÉSIQUE ET GÉOPHYSIQUE INTERNATIONALE

Association Internationale
d'Hydrologie Scientifique

ASSEMBLÉE GÉNÉRALE D'OSLO
19-28 août 1948

TOME II

Procès-verbaux des séances

Travaux de la Commission de la Neige et des Glaciers

Imprimerie Ceuterick
66, rue Vital Decoster
LOUVAIN (Belgique)

COMMISSION OF SNOW AND GLACIERS

COMPTES-RENDUS DES SEANCES

First Ordinary meeting

Friday, August 20th, 9.30 a.m.

In the Chair : Dr. Church (president).

Present :

Argentina : Marchetti.
Belgique : Tison
Canada : Baird, Klein, Legget.
Czechoslovakia : Smetana.
Danmark : Bretting, Ebert, Lyshede, Nielsen.
Finland : Jurva, Korhonen, Renqvist, Siren.
France : Frolov, le Strat, Vibert.
Great Britain : McClean, Manley, Seligman, Wordie.
Hungary : Mosonyi.
Island : Thorarinsson.
Italien : Frosini, Tonini.
Nederland : Thijsse.
Norge : Fægri, Jakhell, Klæboe.
South Africa : Kent.
Sverige : Melin, Slettemark.
Suisse : Hoeck, Mörikofer, de Quervain, Mercanton.
U.S.A. : Beij, Bernard, Sayre.

The President gave an address : Reconstruction or progress.

Fægri presented a report by Schell : Remarks on the distribution of icebergs in the northern hemisphere with special reference to South of Newfoundland and possibility related precipitation. With the report were submitted some shorter papers by Schell and Ferraz.

Korhonen presented a general report on snow cover in relation to runoff. With the report were submitted papers by Korhonen, Jurva and Kaitera.

McClean presented a report on Flooding due to the melting of snow and ice with special reference to the Dee, Aberdeenshire. The report was followed by remarks by Bernard, Fægri, Melin and Slettemark.

Fægri asked for further information with regard to the vertical distribution of rain-gauges. McClean answered that they were all situated in the valley. Melin : We have in Sweden made the same experiences as Bernard and his co-workers in USA. We have found that we get rather good predictions of spring floods from the precipitation stations in the valleys because of the good correlation between these values and the runoff. It is therefore not necessary to measure the precipitation in the high mountains. However, this method of prediction can only be used where there is no melting during the winter.

Fægri presented a report by Gerdel on the American work on physical changes in snow-cover leading to runoff. The report was followed by remarks from Fægri, Legget and Seligman.

Legget asked about temperature gradients in the snow described in the report since the available in the rain (if above 32° F) would have the effect of melting the ice layer, the degree of melting depending in the initial temperature of the ice layer.

Seligman pointed out that ochre strewed on the surface of the Aletsch glacier showed that melt water (as opposed to rain) did not sink to a lower level than 80 cm after 5 weeks although it had travelled a distance of 30 m.

Fægri stressed the possible difference in behaviour between the non-soluble ochre and the soluble fuchsine used by Gerdel.

de Quervain presented a report on the disintegration of the alpine snow-cover. The report was followed by remarks from Legget and the author.

Legget asked if the diagram of temperature showed actual temperatures or temperatures calculated from conductivities. Similar patterns of temperature variations were being obtained in experiments with soil alone did not control the temperature variations. It appears that another very important factor is the movement of water vapor in the voids of the soil. If this is true of soil, a relatively dense material, it would appear probable that the same should be true for snow, a relatively porous substance. *de Quervain* answered that we are not able to calculate temperatures of the snow, because our knowledge about (a) heat conduction in the snow, and (b) heat economy are not of sufficient accuracy. Thus, in the example given, the temperatures are measured.

Mercanton presented a report by Streiff-Becker : Der Wasserabfluss in einen Gletschertal.

The meeting adjourned.

* * *

Second ordinary meeting

Friday, August 20th, 2.30 p.m.

In the Chair : Dr. Mercanton.

Seligman presented a report on the growth of the glacier crystal.

Mercanton presented a report by Renaud : Contribution à l'étude du grain de glacier.

Ahlmann presented a report on recent Swedish work on glacier ice crystal structure. The report was followed by remarks from Seligman.

Seligman said : It is very gratifying to me that this paper gives so much support to the facts I have tried to ascertain in other parts of the world. I am particularly interested in prof. Ahlmann's finding of small crystals in a cliff below an ice fall — the shape and size of these crystals is strikingly similar to the ones I have just described to you in the Upper Grindelwald glacier. I could only wish we knew the size of the crystals *above* the ice fall in this Kebnekaise glacier.

May I press for a uniform method of presentation of the results so that they may be compared with other results. I have been promised data on crystal size from many parts of the world, and much labour will be saved by uniformity. I think my presentation of the crystal grouping in a single sample by histograms is easier to examine and more complete than prof. Ahlmann's graphs. I also strongly recommend expression of crystal size by diameter arranged logarithmically rather than areas arranged somewhat at random.

In preparing an ice surface for recording I have found a proposal casually made to me by Dr. Mercanton very useful. By laying a flat aluminium plate on the ice it absorbs enough heat from the sun (especially if the upper surface is painted black) to melt a glass-smooth surface in a short time. On cold days the plate has to be warmed with a « Meta » iron or blow lamp, or a heavy stone placed on it gives good results, although not so quickly.

Faegri presented a report by Fisher : The pressure melting point of ice and the excavation of cirques and valley steps by glaciers. The report was followed by remarks from Seligman.

Seligman said : Mr Joel E Fisher, like myself, is an amateur glaciologist, and his interest in and contributions to glaciology are well-known. Nevertheless I cannot let pass his assumption that the ice at the bergschrund is the deepest part of the glacier — on the contrary it is nearly the shallowest. In the Great Aletsch glacier, for instance, it is the order of 100 m, whereas lower down it is as deep as 700-800 m. There are other dangerous assumptions too and, I fear, a good deal more speculation than observational basis to his deductions.

Faegri presented a report by Heinsheimer : Horizontal age differences in glaciers.

* * *

Third ordinary meeting

Saturday, August 21st, 2.30 p.m.

In the Chair : Dr. Ahlmann.

Present :

Argentina : Marchetti.

Canada : Baird, Klein, Legget.

Finland : Korhonen, Siren.

Great Britain : Manley, Seligman.

Island : Thorarinsson.

Norge : Fægri, Klæboe, Stavøstrand.

Suisse : Hoeck, de Quervain.

Sverige : Melin.

U.S.A. : Church.

Wordie presented a general report on the variations of glaciers throughout the world. With the report were submitted special reports dealing with the following areas : Iceland, Norway, Sweden, and Switzerland.

Fægri presented a report by Field on the variations of Alaskan glaciers.

Fægri presented a report by Spann on glacier recession in Peru.

Fægri presented a report by Oppenheim on deglaciation in Colombia.

Manley presented a report on Glacier variations and climatic fluctuations in Britain.

Mercanton presented a report by Billwiller : Die kurzfristigen Schwankungen der Alpengletscher and ihre Ursachen.

Fægri presented a report on the variations of western Norwegian glaciers during the last 200 years.

The meeting adjourned.

* * *

First extraordinary (business) meeting

Saturday, August 21st, 12 noon.

In the Chair : Dr. Fægri.

Fægri presented a series of proposals made by Dobrowolsky and originally presented to the congress in Edinburgh in 1936, then to the Washington congress in 1939 and repeated now. The following decisions were made by the commission :

1. The term « cryology » was rejected as far as this commission is concerned.

2. As there are already in existence two periodicals dealing with the subject of snow and ice (*Journal of Glaciology*, *Zeitschrift für Gletscherkunde*) this commission is of the opinion that present circumstances do not favour the establishment of a third international periodical dealing with the same themes.

3. With reference to the planned encyclopedia of cryology this commission authorises the President to take such steps as are necessary to give recognition to the work and consider what should be done to salvage those parts of the work which should be preserved for the future.

4. It is considered that the International Commission of Snow and Glaciers functions and will in the future function as an international institute in the science of snow and ice.

5. Dobrowolsky's proposals with regard to the collaboration between hydrological and meteorologica services were submitted to the Association for consideration by that body.

The meeting adjourned.

* * *

Second extraordinary meeting

Monday, August 23rd, 2.30 p.m.

In the Chair : Dr. Korhonen.

Fægri presented a report by Hoover on the effects of chinook (foehn) winds on snow cover and runoff. This report was followed by a comment from Church.

Church said : The hydrologic interest in Mr. Hoover's paper is the character of snow loss, whether by evaporation or melting. This might be determined by a snow-evaporation pan set in the snow cover and by a drip pan to check it. But the snow should be sufficiently deep to protect the orifice of the drip pan from freezing. One must moreover be on the alert to have the evaporation pan loaded when the foehn approaches. A survey of the depth but particularly of the water equivalent of the snow cover should likewise be made both before and afterwards. Unfortunately the snow cover of the foehn country of the plains east of the Rocky Mountains of Canada and adjoining United States is shallow and subject to sudden melting.

Mercanton presented a report by Haefeli and Kasser : Observations in the firn and ablation region of the Great Aletsch Glacier.

Bernard presented a report by Rhodes and Wilson : The cooperative snow investigations program : its objectives and operations.

Banerji presented a report on Himalayan snowfall and snow-cover observations made by the India meteorological Department.

Read by title : *Heinsheimer* : On the relation of precipitation, accumulation and melting of snow to the stream in the San Juan River, Argentina.

The meeting adjourned.

* * *

Third extraordinary meeting

Tuesday, August 24th, 2.30 p.m.

In the Chair : Mr. Seligman.

Legget presented a report by Meek on glacier observations in the Canadian Cordilleras.

Fægri presented a report by Odell on the glacial features of the Lloyd George Mountains, British Columbia.

Alhmann presented a report on Recent glaciological investigations in Sweden and also plans for the Norwegian-British-Swedish Antarctic Expedition 1949-51.

Church presented a report by himself and Heinsheimer on glacier penitentes in the Andes of Argentina.

Baird presented a report on recent glaciological investigations in the Mt. Vancouver area.

Marchetti presented a report : Studies on snow and glaciers in the Argentine Republic (read by *Fægri*).

The meeting adjourned.

* * *

Fourth extraordinary meeting (joint meeting with comm. of Limnology).

Wednesday, August 25th, 4 p.m.

In the Chair : Dr. Renqvist (President, Commission of Limnology).

Fægri presented a report by Kaitera : Keeping the water from freezing by means of compressed air.

Devik presented a report on supercooling and ice formation in open water.

Faegri presented a report by Lambor : La genèse de la glace flottante. The following proposals made by the author were adopted by the meeting :

1. L'association internationale d'hydrologie scientifique reconnaît la nécessité d'attirer l'attention des savants hydrologues sur l'importance de l'élaboration de la prognose des phénomènes glaciaux sur les cours d'eau.

2. L'association internationale d'hydrologie scientifique reconnaît qu'il est bien à souhait d'indiquer dans les publications hydrographiques la température de l'eau par jour, du moins pour plusieurs emplacements du bassin d'un fleuve. Le mesurage de la température de l'eau pour la chaleur de l'eau au-dessous de 2° C doit être rendu plus exact jusqu'à 0.01° C. En outre, il est nécessaire d'augmenter l'exactitude de l'indication des phénomènes glaciaux en ce qui concerne le temps de l'apparition ainsi que la sorte du phénomène, en introduisant des symboles uniformes. De même il faut indiquer le commencement d'un phénomène avec l'exactitude jusqu'à une heure.

The meeting adjourned.

* * *

Fifth extraordinary meeting

Thursday, August 26th, 2.30 p.m.

In the Chair : Dr. Wordie.

Present :

Canada : Baird, Klein.

Finland : Korhonen, Siren.

Great Britain : Croft, Manley, Seligman.

Italia : Rossolasco, Tonini.

Norge : Fægri.

Suisse : Mörikofer, de Quervain.

Sverige : Ahlmann, Melin.

U.S.A. : Church.

Church presented the following reports : (a) Standardising methods of snow cover measurements. (b) The evolution of snow-melt by dyes and drippan.

Mörikofer presented a report on the dependence of snow cover on altitude in the Alps.

Seligman presented a report on the progress of glaciological research in Great Britain.

Faegri presented reports by Schaeffer, Lundquist and Smith Johansen on the classification of snow and (by title only) a report by Schaeffer on the formation of ice crystals in the laboratory and the atmosphere. In connection with the reports *de Quervain* gave an outline of the snow classification system used in Switzerland, and *Klein* of that in use in Canada. It was resolved that a committee should be appointed to report as soon as possible on the possibilities of standardising an international snow classification system. The President was authorised to elect the members of the Committee. (The following persons have later been elected by the President : Klein, de Quervain, Schaeffer).

The meeting adjourned.

* * *

Sixth extraordinary meeting

Friday, August 27th. 9.30 a.m.

In the Chair : Dr. Faegri (elections), dr. Church (reports).

Present :

Canada : Davies, Klein, Legget.

Great Britain : Manley, Rankine.

Island : Thorarinsson.

Suisse : Mercanton, de Quervain.

Sverige : Ahlmann, Melin.

Elections :

President : Ahlmann (Sweden)
Vice-president : Wordie (Gt. Britain)
Secretary : Baird (Canada)
All elections were unanimous.

Seligman proposed that *dr. Church* be elected permanent honorary member of the Commission in recognition of his great services.

Wordie proposed that *dr. Mercanton* be elected honorary member of the Commission.

Both proposals were carried unanimously.

With the consent of all members present it was decided that the name of the Commission should in the future be : **The Commission of Snow and Ice.**

The executive body of the Commission was authorised to take such steps as it might find necessary for the establishment of an international snow year.

Legget presented a report : Canadian interest in snow and ice research.

Klein presented a report : Canadian survey of physical characteristics of snow cover.

Mercanton presented a report : Les sondages sismiques de la Commission helvétique des glaciers.

Read by title : Sur l'activité de l'institut fédéral pour l'étude de la neige et des avalanches au Weissfluhjoch sur Davos, Suisse, by *Bucher*.

At 11.00 a.m. the meeting continued in the Victoria Cinema where the following films were projected :

Les sondages sismiques de la Commission helvétique des glaciers (presented by *Mercanton*)

The advance of the Grindelwald glacier (presented by *Mercanton*).

Avalanche research station Weissfluhjoch (presented by *Bucher*).

Production of ice crystals in a supercooled cloud (presented by *Schaeffer*).

« Snow crystals » (presented by *Nakaya*).

« Frost flower » (presented by *Nakaya*).

« Snow harvest » (presented by *Work*).

The films had collected a large audience, both from the Association of hydrology and from other associations, especially from the Association of meteorology, the members of which were specially invited.

COMMISSION DE NEIGES ET GLACES

I) ADDRESS OF DR CHURCH — PRESIDENT

RECONSTRUCTION OR PROGRESS ?

The field

The international Commission on Snow and Glaciers is old in years tho not in meetings. The earlier Commission of Glaciers was the child of an earnest desire to accumulate continuous measurements of the advance and retreat of the glaciers and was accepted as a near-hydrologic science by the Association of Hydrology.

The younger Commission of Snow was fostered by the American Geophysical Union because its own Committee on Snow could not accept members from far Australia. With the consent of the Association on Meteorology, the Association of Hydrology again consented to extend its support to a Commission on Snow, but with the distinct understanding that its function would be the development of snow surveying and forecasting of water supplies. The organization of the Commission was assigned to me as American proponent of snow-surveying, but with Prof. Dr. Mercanton of Switzerland and Prof. Eredia of Italy as advisors and guardians.

This action was taken at Lisbon in 1933, but at the meeting at Edinburgh in 1936 the response throughout the world had been so spontaneous and abundant that a volume of transactions of more than 800 pages resulted, including glaciers.

Nearly the entire field of snow and ice science was represented, such as : Precipitation, Snow Cover, Forecasting Streamflow, Runoff, Lake and River Ice, Transport, Physics of Snow and Ice, Avalanches, Glaciers, Arctic and Antarctic Snow and Ice, Reports of Recent Expeditions, Apparatus, Agenda.

The forum was large and individual as well as national, but national groups also sought self-expression. Membership was assumed as permanent and was considered a mark of international confidence. Twenty-seven nations joined in the effort. Nonadhering nations were included and made important contributions. Latvia provided Peter Stakle, Secretary of the Commission and co-editor of Bulletin 23.

The younger glaciologists who sought the physical approach to the study of glacier phenomena were eagerly welcomed by the Commission of Snow, but felt the awkwardness of the invisible but formal boundary between the alimentation and dissipation zones of the glacier. They realized the essential identity of seasonal and perennial snow and ice and Dr. Dobrowolsky had already long urged the compilation of an Encyclopedia on Cryology or Science of Cold.

At the Washington meeting in 1939 the Commission on Glaciers generously acquiesced in the general desire to unify the two commissions. In return an even broader foundation for the Permanent Measurements of Glaciers throughout the world was established in the joint Commission of Snow and Glaciers.

Special projects were initiated for (1) Standard maps of snow-cover and ice-cover, (2) Standard classification of types of snow and snow-cover, and (3) A suitable classification for the International Bibliography of Snow and Ice already begun, and (4) Standardizing methods of snow-surveying and forecasting runoff from snow-cover. Ice patrol and permafrost were added to the earlier interests and snow and ice engineering have recently been suggested by Canada, Soviet Russia and Switzerland.

The Future

The original purpose for which the Commission was organized is now being fulfilled. Snow-survey and streamflow forecast systems have been established in the snowed areas of North America and in the Himalaya and Southern Andes. International cooperation has long been practiced by Canada and the United States and is being established by India and the Himalaya states and by Argentina and Chile who must share both water and power as well as the snow-surveys.

Iran, Spain, Korea, Norway, Sweden, Switzerland, wherever water storage is insufficient for irrigation, power or seasonal stream uses, are fields for the Commission's study. Iran and Spain have already made inquiries and in Norway the problem has been assigned to a commission for investigation, and snow measurements have been begun in Switzerland.

The economic value of glaciers as providers of water has been much in the public

and scientific mind in India and South America, and glacier water is in use in Switzerland and Norway. As compared with seasonal snow, this source is temporary and may vanish. At best it provides only late water because of the overlying cover of snow and its relatively small proportion in the seasonal flow. All angles of this problem should be carefully studied.

The time, however, has come to give formal recognition to the modern science of snow and ice with its multiple relationships and human services. The chart in Bull. 23, p. 5, of the first meeting of the Commission of Snow is still our guide. « Sports » and « permafrost » might be added because the terms are more modern.

The projects in snow-surveying and streamflow forecasting can now carry on. Economic accuracy has now been attained. Supervision of flood storage has been inaugurated. The characteristics of snow-melt are now under investigation. The original purpose of the Commission has become only a memory.

The world study of snow and ice is at hand. The technical study of these elements is being pushed to the ends of the earth. A Swedish-Norwegian-British Expedition to Queen Maude Land is under way. Argentina will inaugurate a study of snow in the South Orkades and Graham Land and Denmark will make observations in Peary Land. The snow laboratories of the mountains of North America and Europe are natural centers for this work, and high mountain observatories, as on Mount Washington, afford abundant icing. The study of the origin of snow crystals and their precipitation is advancing rapidly at Hokkaido University, Japan, and the General Electric Company of North America. Russia lost its Arnold-Alabiev in the starvationsiege of Leningrad.

But Dr. Kalitin of albedo fame during our visit at the CCXX Anniversary of the Soviet Academy of Sciences in 1945 proposed a world snow-year during which all snow countries of the earth would make daily observations of the physical characteristics of the snow-fall and snow-cover and share their reports. Could not these studies be synchronised with the South Polar trip in prospect ? except that the Northern end Southern seasons are chronologically opposite, as one realizes who flies quickly from one end of the earth to the other. The key stations are ready. Special snow-observations stations to supplement them need only be selected by the various countries. Methods already proposed will be presented at the sessions. The Commission should foster and conduct the work.

Personnel

War may interrupt or even stop progress. It may even accelerate it. In the Americas and Asia the work has gone rapidly forward. The « Reorganization Letter » sent to « Surviving Colleagues » in February, a year ago, has revealed 62 previous members who answered the rolleall, and many had reports to make. There were 22 more whose activity is known, or a total of 84 who are eager to maintain membership.

The + marks the names of 22 not including those who died previous to the Personnel Revision of December 1, 1946. Among these were most of the Soviet Russian Group and their leading spirit, Boris Weinberg, who wrote me precious love messages on a post card.

Of new members nominated from various countries of the world there are 62, representing ambition and youth.

Our progress lies before us. Our membership is ready. We need no reorganization except freedom to attain our natural growth. The Commissions are groups drawn together by mutual scientific interest and permanent in character as compared with temporary committees. Their specialities within themselves may be varied and changing. They will become national groups or even inspire separate national organizations and movements. Our membership gives us buoyancy and momentum. The associations joined as an International Union give us world perspective.

Thru death we have lost heavily in members and leaders, but we have gained immeasurably in inspiration. The following notes and messages are typical :

« Professor Ulrich died 20th October, 1941, manfully resisting the Germans.

He was one of the more than eighty Czech men of science who became victims. »

A year ago our President Lütseh, a monk in gentleness, but a crusader in action, died in the mountains at his work. In early 1945 he had written to me :

« With great interest I have noted your report of the activity of the Snow and Groundwater Commissions. This demonstrates that high activity can be

attained if the presidents of the individual commissions (i.e., the commissions themselves) are assured the greatest possible independence.

« I myself feel strong and well, I work daily 6 to 10 hours on the completion of my life-work. This work keeps me young and fresh.

« Every best wish : Devotedly

 Your O. Lütshg

« Best wishes and greetings also to your Colleagues. »

And now at the height of their activity in preparation for the present assembly, my colleague Secretary and Co-president of the Commission, Dr. François E. Matthes, and Dr. Oscar E. Meinzer, President of the Commission on Ground water, have died. Dr. Matthes, after a previous illness, sturdily took over my duties as managing director of the Commission while I lingered on in South America to complete the snow-survey system in the Andes. After he was stricken and struggling for life, Mrs. Matthes took over until the official work could be transferred to Acting Secretary Linsley at Washington, D. C. But she stoutly disclaimed the title of Associate Secretary which she had fully earned as personal aid to her husband.

Dr. Meinzer slept away quietly after a long conference by telephone.

Of those urged to join the Commission because of their wisdom, tho their days of creative activity were over, was Dr. Tutton of the English Group, who died in the dark days of the war. The notice reads, « Will you kindly make a note of the death of my brother in-law, Dr. A. E. H. Tutton, E.R.S. Some years ago. Your notice of books came to hand here... I wonder you didn't see the notices in the various papers at the time of his death, especially the long one in the Times. He is a great loss to the scientific world, as well as to his family. »

Of those still living, but whose hand trembles with infirmity and age is Dr. Ing. Cenek Vorel, former Director of the National Hydrological Institute of Prague who writes,

« I should like to thank you for the Provisional Personnel List of the International Commission of Snow and Glaciers and also for the two Supplemental Letters.

« I have retired, however, several years ago, and my health does not permit me to take effective part in the above Commission, whose member I am.

« I have therefore, passed on the papers you sent me to Eng. F. Kovarik, the present director of the National Hydrological Institute, and asked him to take over my duties.

« I should like, therefore, Mr. President, to be relieved of my membership of this Commission.

 Yours faithfully... »

He should be retained as Emeritus.

And now a younger message from far Japan of scientific industry and happiness that peace has come.

« With much delight I received your letter of Feb. 7th, 1948, sent out from Buenos Aires. I feel it very happy to tell you that I have passed the hard time safely and have been able to continue my work on snow. My experimentation made much progress during those several years, and now all sorts of snow crystals are made in the laboratory at will. Some of them are more beautiful than the natural ones, and the meteorological conditions controlling the formation of various types of crystals are now almost clarified. » — Ukitiro Nakaya.

We may have Prof. Nakaya and his film with us if the spontaneous assistance of American Government and scientific agencies can bring the miracle to pass. If delay occurs, the film is still assured to all countries requesting it.

And now 15 years or a half a generation has passed for our mutual membership in the Commission. Some of us have grown old in the service. Some are in the strength of middle age. Some of you have come with the idealism of youth. To have known you all personally by face or letter has made my life rich indeed. It is now my turn to step aside and yield the presidency to other colleagues.

The future belongs to you.

 s/s J. E. Church



II. — Rapports relatifs aux :
ICEBERGS

1) **ON THE DISTRIBUTION OF ICEBERGS IN THE
NORTHERN HEMISPHERE WITH
SPECIAL REFERENCE TO SOUTH OF NEWFOUNDLAND ***

by

I. I. SCHELL

Woods Hole Oceanographic Institution

(Abstract)

Systematic compilations of the iceberg count off Newfoundland covering a period of 69 years of record (1880-1948) and observations of bergs off the East Greenland coast and in the Arctic north of Siberia appear to indicate a somewhat greater number of drifting bergs in recent years in north polar waters.

The apparent increase in loose bergs is probably due to the widespread warming up in recent years, making possible their release from ice locked fjords and bays where in some instances they have been held for a long time.

The presumed freeing of bergs in Baffin Bay and Davis Strait by factors not directly related to the pressures and temperatures in the North Atlantic which were previously indicated by Smith and Groissmayr to affect their distribution off Newfoundland, has complicated the prediction of the severity of the iceberg season in that area.

The role of the circulation also to the west of Greenland, recently considered by Walker, must be further studied.

**1. An apparent Increase in Loose Bergs Observed in Recent Years
off Newfoundland and Elsewhere in North Polar Waters**

Available observations of drifting icebergs off Newfoundland, off the East Greenland coast, and in the Arctic North of Siberia seem to indicate an increase in their number. Bergs that have apparently been ice locked in the fjords of East Peary Land since 1892 were freed (Koch, 1945) shortly before 1938. Many icebergs have also been observed (Zubov, 1945) by recent Russian expeditions in the eastern Arctic. Also hundreds of bergs were observed from the air in 1943 north of Franz Joseph Land in latitude 84° N and in 1945 north of Northern Land (Stefansson, 1947).

The average severity of the iceberg season south of Newfoundland appears to have slightly increased in recent years despite two phenomenally light seasons in succession, 1940 and 1941. The average value of the berg count on scale of 10 for the first third (1880-1920) of the available 69-year period of observations 1880-1948, is 4.6, while for the last two thirds (1903-1948) it is 5.0**.

While it is possible that the comparatively fewer bergs observed off Newfoundland during 1880-1902 are due to fewer ships and perhaps less systematic observations in these waters in that period the effect of this is probably more than balanced by some unobserved bergs during the first period of disturbed world conditions affecting the operations of the International Ice Patrol. (During World War II the duties of the I. I. P. were taken over by other vessels.)

The apparent increase in the number of bergs off Newfoundland and their appearance in large numbers off the upper reaches of the East Greenland coast, north of Franz Joseph and Northern Land and in neighboring Arctic waters if real, would not be inconsistent with the observed general warming up in recent decades in the high latitudes of the northern hemispheres which would have made possible their release from ice locked fjords, bays, etc.

(*) Read at the meetings of the International Commission of Snow and Glaciers in Oslo in 1948.

(**) Based on figures published by the U. S. Coast Guard.

Références

- KOCH, L. (1945), The East Greenland Ice.
Meddel. om Grønland, Bd. 130, Nr. 3, pp. 1-374 (109)
København.
- STEFANSSON, E. (1947), The eleventh visit to the North Pole
Geogr. Rev., v. 37, No. 1, pp. 153-4. New York, N. Y.
- ZUBOV, N. N. (1945), Ldy Arktiki (Ice of the arctic).
North Sea Route Publications, pp. 1-360 (102-104).
Moscow.

2. On Foreshadowing the Severity of the Iceberg Season Off Newfoundland

The need for studying additional factors which determine the severity of the iceberg season south of Newfoundland to which attention was previously called by Smith and by Schell (1940) is emphasized in a recent paper by Walker (1947).

Walker's examination of the pressure variations in the northern North Atlantic, preceding the iceberg season, indicates that beginning shortly after the turn of the century the pressures at Iceland, southernmost Greenland, eastern Labrador, and southern Norway, which form the substance of Smith's formula (1931) for predicting the severity of the iceberg season, were apparently no longer a factor during the period 1906-1938. While it may be significant that in this study Walker considered the pressures for not quite the same season as was done by Smith (*ibid.*), possibly making for only a diminished control of iceberg conditions off Newfoundland by the circulation in the northern North Atlantic, the real reason for the apparent breakdown in the control probably lies elsewhere.

In an attempt to account for the observed variations in the iceberg count off Newfoundland during the latter half of the period which he analyzed, Walker (*ibid.*) finds an apparent relationship with the pressure at Point Barrow during the preceding October-November, high pressure tending to precede a light iceberg season and low pressure a heavy iceberg season off Newfoundland the following spring and early summer. He suggests that the control of Point Barrow pressure lies through the winds which drive the icebergs originating west of Baffin Bay through Lancaster Sound and Jones Sound into the Labrador Current and assumes that low pressure at Point Barrow intensifies an existing south to north pressure fall over Alaska which would be associated with stronger westerly winds over the entire region from Point Barrow to Baffin Bay. The interval, October-November to April-May, when the iceberg season is at its heaviest off Newfoundland, presumably corresponds very roughly to the time estimates for the icebergs from west of Baffin Bay to arrive off Newfoundland.

Two difficulties seem inherent in the mechanism suggested by Walker. While the prevailing winds west of Baffin Bay are from the northwest as part of the circulation associated with the Baffin Bay low, the winds at Point Barrow during October and November as well as the rest of the cold season are generally from the northeast or east in conformance with the circulation in the northern section of the Aleutian low and in the southern part of the Arctic high farther north, in either of which Point Barrow may lie, depending on the positions of the two systems. The other difficulty in Walker's reasoning is in assuming an appreciable number of icebergs drifting into Baffin Bay from the west. According to Smith (*ibid.*), this number is quite small and in comparison with those arriving from the east side of Baffin Bay, entirely negligible.

The existence of winds from a northeasterly direction at Point Barrow during October and November does not itself exclude the pressure in that region from being a factor in the severity of the iceberg season off Newfoundland but it indicates that the mechanism suggested by Walker is too simple. The apparent role of Point Barrow pressure in the control of iceberg conditions in Baffin Bay and Davis Strait and subsequently off Newfoundland may rather be through its relation to the pressure fall west of Baffin Bay to Alaska and consequently to the variations in the northwesterly winds that affect the iceberg supply in Baffin Bay. Also the smallness in the number of bergs available from west of Baffin Bay for ingress into the Labrador current is no serious obstacle to Walker's thesis that conditions over northern North America are a factor in the iceberg count off Newfoundland, for, as is well known, icebergs from West Greenland find their way in large numbers to the

west side of Baffin Bay and Davis Strait from where they are subsequently released after a more or less indefinite interval of time.

It should also be noted that following Smith's derivation of a formula for predicting the severity of the iceberg season off Newfoundland based on pressures only (*ibid.*), Groissmayr (1939) derived one based mainly on temperature variations in the North Atlantic. According to a test made by Schell (1940), Groissmayr's formula gave somewhat better agreement between the computed and actual berg count for the period examined than the earlier derived formula, especially when considered for cases when the computed deviations exceeded $0.841 k. s$, where $k = \sqrt{1 - r^2}$, r being the correlation coefficient, and s the standard deviation of the iceberg count.

Finally, it was indicated by Soule (1940) that years with few bergs off Newfoundland appear to coincide with lower than normal temperatures and smaller volumes of flow of the Labrador Current. The apparent relationship between the iceberg count and the heat transfer of the Labrador Current is tentatively explained by Soule to arise from the winds over the Davis Strait and Baffin Bay region. With increasing southerly winds the volume of relatively warm water that branches off from the northward flowing West Greenland Current to cross Davis Strait subsequently to merge with the colder Labrador Current is diminished. The stronger southerly winds would at the same time tend to impede the descent of some icebergs carried southward along the west side of Baffin Bay and Davis Strait.

In addition to the desirability of making a broader study of the oceanographic and meteorological features of the circulation that are linked with iceberg conditions off Newfoundland, the availability of bergs in Davis Strait and Baffin Bay and along the Labrador coast must also be considered for possible foreshadowing the severity of the iceberg season farther south each spring and early summer.

REFERENCES

- GROISSMAYR, F. B. (1939), Schwere und leichte Eisjahre bei Neufundland und das Vorwetter.
Ann. Hydr. v. 67, pp. 26-30.
- SHELL, I. I. (1940), Foreshadowing the Severity of the Iceberg Season South of Newfoundland.
Bull. Am. Met. Soc., v. 21, pp. 7-10.
- SMITH, E. H. (1931), The Marion Expedition to Davis Strait and Baffin Bay, 1928.
U. S. Coast Guard Bull. No. 19, Pt. 3, pp. 1-221.
- SMITH, E. H. (1940), Recent movements of No. Atlantic ice and a proposed Coast Guard experiment in the West Greenland glacier.
Trans. Am. Geophys. Un., Wash., D. C., v. 21, Pt. II, pp. 668-672.
- SOULE, F. M. (1941), Physical Oceanography — the Grand Banks region and the Labrador Sea in 1940.
U. S. Coast Guard Bull. No. 30, pp. 36-56.
- WALKER, G. T. (1947), Arctic conditions and world weather.
Quart. J. R. Met. Soc., London, v. 73, pp. 226-256.

2) NORTHERN SEA ICE AND RAINFALL IN S. E. BRAZIL

by

J. de SAMPAIO FERRAZ

In our studies of the larger rainfall variations in S. E. Brazil, particularly over the southern section of the S. Paulo plateau where typical cold fronts are frequently developed especially in the wet season, we have recognized in several instances major factors concurring for abnormal precipitation, excessive or otherwise.

One of these major factors, we believe, is embodied in the massive southward extension of arctic conditions over northern Europe or even farther south, and in the observations of arctic sea ice in lower latitudes.

Now, severe winters in the greater part of Europe coincide often with excessive rainfall in S. E. Brazil, a meeting ground of the southern hemisphere's cold and warm air-masses. Amid examples of this coincidence we may recall those of 1870-71, 1928-29, and 1946-47.

The only series of observations of northern sea ice which we have compared with our larger variations of rainfall is the one previously compiled by Smith (1931), that is, the count of icebergs off Newfoundland. For what concerns our studies, there are several instances in which a high number of icebergs corresponds to excessive rainfall in S. E. Brazil; and also cases in which our marked dry spells coincide with small counts of bergs off Newfoundland. Typical examples are the very wet season of 1928-29, occurring with over 1300 bergs, and the 1924 dry spell coinciding with only 11 icebergs.

This apparent relation between arctic southward thrusts and our excessive rainfall in a strategic region of cold fronts, seems to indicate an interdependence of the atmospheric circulation of both hemispheres. We presume such trusts are not simultaneous at both poles. In 1928-29, our tropical (continental) air masses were extraordinarily active (more precisely in January and February 1929), putting up a much greater resistance to the usual advance of migrating anticyclones from the southwest. On the other hand, in 1924, when the arctic ice apparently receded, our south polar airmasses, breaking all records, were abnormally active and moved freely towards the equator.

REFERENCE

- SMITH, E. H. (1931), The Marion Expedition to Davis Strait and Baffin Bay, 1928. U. S. Coast Guard Bull. No. 19, Pt. 3, pp. 1-221 (181).

III. — Rapports relatifs aux :
CARACTÉRISTIQUES PHYSIQUES
DE LA COUVERTURE DE NEIGE

RAPPORT GENERAL

de

V. V. KORHONEN

Ce rapport général présente des résumés des rapports nationaux suivants : Grande Bretagne (Mc Clean), U.S.A. (Forrest L. Rhodes et W. Wilson, D. F. Jones), Canada (G. Klein), Indes (Banerji), Pakistan (Gorrie), Suède (H. Ahlmann), Suisse (de Quervain), Finlande (Jurva, Kaitera et Korhonen).

Monsieur le Président, Messieurs

Des rapports sur les recherches de la couverture de neige sont parvenus des pays suivants : L'Angleterre, l'Argentine, les Etats-Unis d'Amérique du Nord, le Canada, les Indes, la Suède, la Suisse et la Finlande. Leur contenu principal, en considérant surtout la manière suivant laquelle on a organisé l'étude des neiges dans les différents pays, est le suivant.

Un rapport de W. N. Mc. CLEAN, *Flooding due to melting of snow and ice* contient une information et des exemples de ce que la quantité de l'eau coulante produite par la neige fondante est plus grande que celle démontrée par les pluviomètres à l'égard d'un certain flot d'eau causé par la fonte de la neige.

La période froide de janvier jusqu'à mars 1947 causa des inondations sérieuses du Great Ouse, du Trent et du Lee. On a calculé la couche de neige; la fonte de la surface de cette masse ne put être absorbée par les espaces de réserve à disponibles et produisit des inondations sérieuses.

Le compte rendu des recherches étendues entamées récemment aux Etats-Unis d'Amérique sur la composition et la physique de la couverture de neige a été publié dans deux dissertations intitulées toutes deux :

« Le programme de recherches coopératives sur la neige : ses objets et ses procédés », — qui ont été rédigées par le Directeur du programme *Forrest L. Rhodes* et *W. Wilson* et Colonel, l'Ingénieur de division de *C.E. Dwight F. Jones*. Celui-là décrit le programme en traits généraux et celui-ci considère les détails.

En reconnaissant l'influence de la neige sur le cours anticipé des eaux fluviales et sur le projet, le dessin et l'action du contrôle de l'inondation et des projets à fins multiples dans les régions montagneuses des Etats-Unis, un programme étendu de recherches sur la neige a été entrepris en 1945 à l'initiative du Bureau Météorologique et de la Division des travaux civils du corps des ingénieurs, sur la base d'une coopération, afin de

A. déterminer une méthode pratique et sûre pour évaluer l'augmentation maximum de la hauteur des eaux dans le fleuve puisse être causée soit par la fonte de la neige soit par l'ensemble de la fonte de neige et de la pluie.

B. développer une méthode pratique et sûre pour qu'on puisse prédire l'augmentation de hauteur des eaux dans le fleuve, l'inondation y comprise, pendant les saisons et les périodes plus courtes, causée par la fonte de la neige ou ensemble par la fonte de neige et de la pluie.

C. accroître nos connaissances fondamentales des caractères hydrodynamiques et thermodynamiques de la neige..

D. accroître nos connaissances des phénomènes météorologiques, climatologiques et hydrologiques pour trouver comment ils influent sur les cas A, B, et C ci-dessus.

Afin qu'on puisse procéder à ces recherches, trois régions de recherches ont été formées qui représentent des différents facteurs météorologiques et territoriaux. Dans les facteurs météorologiques la nature des précipitations a été considérée comme suit :

- 1) la pluie et la neige avec prédominance de la pluie
- 2) la pluie et la neige avec prédominance de la neige et
- 3) la neige

auxquels les régions suivantes de recherches correspondent comme suit :

1) Willamette Basin Snow Laboratory, un bassin de dérivation d'une aire de 11 miles carrés, situé dans la Cascade Rouge à 40 miles à l'est d'Eugène, Orégon, 1947;

2) Central Sierra Snow Laboratory, un bassin de dérivation d'une aire de 4 milles carrés, situé dans la Sierra Nevada près de Donner-Summit, California, 1945;

3) Upper Columbia Snow Laboratory, un bassin de dérivation d'une aire de 20 milles carrés, situé dans les Northern Rocky Mountains près de Glacier Park Montana, 1945.

Le programme étendu d'observations et de recherches comprend les points principaux suivants :

Radiation directe et reflectée du soleil et du ciel.

Profils de la température de la neige et du sol.

Profils de l'humidité du sol.

Température, humidité et vitesse et direction du vent à la hauteur de 30 à 50 pieds de la terre.

Présentation des pluviomètres de différents types.

L'épaisseur de la neige, valeur équivalente de l'eau et la densité.

Photographies aériennes faites verticalement de la neige 5 fois pendant la saison de fonte.

Photographies en couleur montrant l'humidité dans la neige.

Pénétration de la radiation dans la neige.

Rétention de l'eau et pourcentage d'eau liquide de la neige.

Classification des caractères physiques de la neige.

Un rapport canadien : GEO KLEIN, *Canadian survey of physical characteristics of Snow-cover*.

Cet écrit concerne une étude des couches de neige entreprise au Canada pendant les deux derniers hivers. La méthode, les instruments et les résultats du travail du premier hiver sont présentés.

Le but de cette recherche était de rassembler des données spéciales sur les couches de neige dans différentes zones représentatives données qui pourraient être appliquées à de nombreux problèmes hivernaux, spécialement à ceux dans lesquels l'état de la neige est d'importance. Cette recherche devait aussi former la base d'études futures sur les propriétés de différentes formes de neige tombée.

Pour faciliter la comparaison des résultats de différentes stations, il fut nécessaire de limiter la recherche aux « stations exposées » avec une surface d'essai plate non protégée et aux « stations protégées ». Dans les essais de 1947, seules les stations exposées furent prises en considération.

Des mesurages hebdomadaires furent faits dans chaque couche de neige distincte. Le mesurage comprenait : la température, le poids spécifique, la dureté, la grandeur moyenne et la forme moyenne des grains de neige ainsi que le pourcent du contenu de l'eau libre et la température de l'air, la profondeur totale de la couche de neige et les profondeurs des limites des couches. Des observations nécessaires concernant les caractéristiques du temps comme la température de l'air, la quantité de vent et de soleil ainsi que la qualité et la quantité de la nouvelle précipitation furent faites tous les jours.

S. K. BANERJI, Director-General of Observatories, Meteorological Office, Lodi-Road, New Delhi (India) : Observations of Himalayan snowfall and Snow-cover made by the India Meteorological Department.

L'importance de la chute de neige et de la couverture de neige sur les montagnes au nord des Indes, notamment au cours des mois d'hiver, pour une propre compréhension de la météorologie des Indes a été reconnue dès que l'organisation météorologique des Indes a été établie en 1875 et depuis ce temps des efforts ont été faits pour recueillir des notions régulièrement. Des données qui ont été publiées dans l'annuaire « India Weather Review » ont compris ce qui suit :

1. Le nombre des jours durant lesquels la neige est tombée et une estimation sommaire de la quantité de neige tombée pendant un mois,

2. L'épaisseur de la couverture de neige sur certains postes et stations à la fin du mois.

3. L'altitude la plus basse que la chute a atteint pendant le mois et

4. La comparaison relative à la qualité entre la chute de neige pendant le mois et celui d'un an normal,

De plus, l'India Meteorological Department a entretenu 15 stations météorologiques en Himalaya où des observateurs entraînés recueillent des notions de la pluie et de la chute de neige comme aussi d'autres éléments météorologiques. Entre les résultats de recherches sur la neige doit être mentionnée l'influence de l'abondance de la neige sur le mousson d'été et sur la qualité d'eau du fleuve Indus.

R. M. GORRIE, *Report of Pakistan to international commission on snow and glaciers*

The partition of Pakistan from India in August 1947 was made so precipitately that practically no scientific or official records of any value have been transferred to Pakistan, and most private libraries have been lost including my own copy of your Edinburgh Proceedings which has been kept by the Indian Central Waterways and Inland Navigation Committee in New Delhi. The Pakistan Secretariat considers the matter is not of sufficient interest to establish a CWINC for Pakistan, so little progress is likely to be made if we rely upon official backing.

Our main interest must be in your Question 2, Physical changes in snow cover as affecting run-off and floods but measurement or survey of our Snow catchment is not practicable at least until the Kashmir war ceases and that country is reopened for ordinary travel in the mountain catchments of the Indus, the Jhelum and the Chenab, which constitute Pakistan waterways.

The Uhl valley in Kangra on which Dr. Church noted some years ago when I asked his help on the question of Snow Survey and the Nadir winter flow affecting hydroelectric potential is now in East Punjab and it is not known whether the extensive plans for erosion control, and afforestation are to be continued or not.

The West Punjab has been very badly handicapped by failure of the 1948 harvest in a large irrigated area due to the East Punjab government cutting off all water supplies from the headworks which as a result of partition are outside the borders of Pakistan. The canal affected is the Upper Bari Doab Canal. An alternative channel is now being dug so as to tap the flood — water of the Ravi River at a point inside the Pakistan — boundary.

« Soil and Water Conservation in the Punjab » written by myself and published price Rs : 5/—/— by the West Punjab Govern Press, Lahore includes an attempt to evaluate the catchment values and flood potentialities of Himalayan rivers and torrents.

Sd/— R. M. Gorrie
Conservator of Forest
Rawalpindi Circle

HANS W. AHLMANN, *Recherches sur l'évolution des couches de neige dans le massif du Kebnekajse (Laponie suédoise)*

Dans le cadre des recherches glaciologiques en cours depuis le Printemps 1945, dans le massif de Kebnekajse, l'étude de l'accumulation et de l'ablation des couches de neige a été entreprise.

Une attention particulière a été consacrée aux études des changements physiques qui transforment la neige d'hiver, froide, caractérisée par une forte et irrégulière stratification thermique (avec des variations de 10 à 15 degrés centigrades par 25 cms), en une neige d'été isothermique — les études sont d'une grande importance pour la compréhension de la cristallisation de la neige, des variations de sa teneur en eau et de son coefficient de conduction calorifique.

Pendant le mois de mai, il semble qu'il y ait assez souvent une forte ablation, mais en général — on peut l'expliquer par l'affaissement de la neige à la suite de sa recristallisation, et aussi, quand il se forme à la surface de l'eau de fusion, par la redépôt de cette eau de fusion sous forme de glace dans les couches sous-jacentes. Un léger ruissellement s'écoulant de la couverture de neige persiste jusqu'à ce que l'onde de froid hivernale ait entièrement disparu (ablation nette) — les petites quantités d'eau qui proviennent du dépouillement des couches de neige encore froides, filtrent à travers des canaux particulièrement perméables, là où l'eau à peine au dessus du point de fusion peut pénétrer à travers la neige à une assez basse température. En même temps que ce ruissellement, il se forme aussi une grande partie des lits de glace qu'on trouve toujours dans la neige de Printemps — ces lits de glace

sont la conséquence de surfaces de discontinuité plus anciennes, par exemple la limite entre deux chutes de neige —.

On a aussi étudié, dans des coupes de puits fréquemment creusés non seulement les variations de la température des cristaux, mais aussi celles du poids spécifique; les variations sont si importantes pendant la période d'ablation (le poids spécifique s'accroît de 0,3 — 0,4, en mars, jusqu'à 0,6 en juin), que l'ablation de la neige doit être déterminée, à des intervalles répétés, par une complète mesure de sa teneur en eau. Pour cela, on creuse un puits vers la neige de l'année précédente — ou vers la glace — et on fait des mesures du poids spécifique tous les 25 cms.

De plus, pour pouvoir déterminer certaines surfaces de référence, on a répandu sur la neige une poudre colorée, à différents époques de la période d'accumulation.

M. DE QUERVAIN. *Ueber den Abbau der alpinen Schneedecke et le résumé : sur la diminution par fonte et évaporation de la couche de neige des Alpes.*

Les dernières années on a fait en Suisse de nombreux mesurages et observations qui se rapportent directement ou indirectement à la diminution de la couche de neige. L'établissement de recherches sur la neige et l'avalanche de neige se trouve au-dessus de Davos, Weissfluhjoch, situé à une hauteur de 2.660 mètres. Il se trouve dans la zone entre les frontières forestières et glacières dont les circonstances neigeuses sont l'objet de la recherche en question.

Il s'agit en premier lieu d'observations, de mesures et de calculs établissant le bilan de la quantité de chaleur d'un champ horizontal de neige avant et pendant la fonte. Entre les questions auxquelles la recherche se rapporte, les suivantes sont les plus importantes :

1. L'échange direct de chaleur entre la neige et l'air. On peut le préciser seulement dans quelques cas simplifiés et idéaux.

2. Le courant de chaleur du sol à la couverture de neige.

3. L'évaporation et la condensation. Plusieurs séries de mesures d'évaporation montrent de grandes différences de quantités évaporées suivant que la surface est au soleil ou à l'ombre.

4. La conductibilité de la chaleur de la neige. Il était intéressant d'établir dans quelle mesure l'air des pores de la neige contribue au transport de la chaleur. La conductibilité de la chaleur dans la neige humide se limite au transport de chaleur par les pores et par l'infiltration de l'eau fondue. Lorsque la surface de la couverture de neige fond, l'eau de fonte est l'agent principal de la transmission de chaleur.

5. La désagrégation de l'eau dans la neige et le retardement dans le courant d'eau.

6. La couverture de neige des Alpes pendant la fonte de neige. L'exposé détaillé sur les événements de fonte de la couverture de neige et sur les conditions qui peuvent amener aux affluences anormales et dangereuses de l'eau.

R. JURVA. *Über die Eisverhältnisse des Baltischen Meeres an den Küsten Finnlands.*

Dans ce travail on a présenté une méthode catégorique envisageant les possibilités d'étudier des phénomènes limités à des régions définies, mais montrant une très grande variabilité par rapport au temps. La méthode a été appliquée aux glaces de la Mer Baltique sur les côtes de la Finlande.

P. KAITERA. *Snow melting in the spring and its effects on the opening of the waterways in Finland.*

L'écrit est un court résumé d'une large recherche. Comme matériaux ont servi les mesurages faits pendant la fonte printanière en 1934-37 de la profondeur de la neige et de la quantité de l'eau, qui ont eu lieu comme mesurage de ligne à différents endroits dans des terrains de différents types.

Sur de petites plaines forestières, dans les forêts d'arbres à feuilles, dans les forêts de conifères espacés et dans les forêts mixtes il y a plus de neige, mais dans les forêts de conifères denses moins que sur les champs découverts. Il fond de la neige au printemps; quand la vitesse de la fonte est la plus grande, au cours de 5 jours environ 45-50 pour cent par jour, en 10 jours environ 70 et en 15 jours environ 85 pour cent de la quantité d'eau la plus grande et ces chiffres relatifs sont dans les différentes parties du pays et dans les types de terrain différents à peu près les mêmes que les submentionnés.

Le soussigné a composé deux écrits : *Report of the researches concerning snow-and ice condition, which have been made in Finland after 1936* et : *Correction des mesurages des chutes de neige faite sur la base de la nature de la place de mesurage.*

Au Congrès d'Edimbourg en 1936 je présentai un rapport concernant les observations faites en Finlande sur la neige et sur la glace et concernant les recherches dans ce domaine; je ne veux donc pas me répéter ici. Le programme des observations est toujours le même. Des changements dignes d'être mentionnés n'ont pas eu lieu excepté en ce qui concerne l'Institut Météorologique Central qui vient d'élargir le programme de ses observations sur les chemins praticables aux traîneaux d'une manière considérable, et qui a commencé à partir du 15 janvier à faire des mesurages de terrain de même sorte que ceux du 15 mars.

Ci-jointe une liste de recherches faites depuis 1936 ainsi que la liste présentée antérieurement comprenant des questions nombreuses concernant les conditions de neige et de glace. Avant tout on a essayé de développer un procédé pour obtenir l'évaluation la plus correcte possible de la quantité d'eau contenue dans la couche de neige et pour mesurer de la manière la plus exacte l'erreur de mesurage de la neige commise avec le pluviomètre. A cet égard on a développé un procédé de correction où la correction de la quantité de neige mesurée par le pluviomètre se base sur la nature des alentours de la place de mesurage.

Concernant la fonte de la couche de neige on a fait des essais au printemps dans le but d'éclaircir la dépendance des inondations de printemps. Vu que les différences de hauteurs en Finlande sont de peu d'importance, excepté dans quelques parties de la Laponie, mais qu'au contraire la différence entre les champs découverts et les différentes sortes de forêts est considérable, on a en premier lieu fait attention à ceux-ci dans ces recherches.

Des recherches spéciales concernant la construction de la couche de neige ont été faites en Finlande en connection avec d'autres recherches concernant la géophysique du pays. Le plus souvent il a alors été question de conditions de chaleur et de gelée du sol du pays et il a été nécessaire d'éclaircir l'importance de la couche de neige, c'est-à-dire la conductibilité de chaleur de la neige à cet égard. On a pris en considération ensemble avec la profondeur de la couche de neige seulement la densité. On n'a observé la structure, les stratifications, la cristallisation, la grandeur des grains de neige etc. qu'à vue d'œil, sans aucun instrument.

Les rapports précédents sur la physique de la couverture de neige se réfèrent en général aux circonstances dans la montagne. Aux circonstances similaires se rapportent aussi plusieurs des recherches antérieures sur la composition intérieure de la couverture de neige se référant en premier lieu à la connaissance des avalanches. Sur les circonstances dans les terrains bas et relativement plats il existe seulement quelques recherches spéciales séparées. Il serait cependant bon si on arrangeait de la même manière qu'aux Etats-Unis et en Suisse des terrains d'essai pour des recherches de ce genre aussi quant aux terrains bas dans les pays où la couverture de neige est considérablement épaisse. Les différences de hauteur étant sans signification l'attention sera prêtée en premier lieu à la différence de qualité de la couverture de neige entre les terrains découverts et les terrains boisés de différents genres. Accomplies dans des circonstances différentes ces recherches apporteraient une clarté plus cohérente que les recherches d'une nature casuelle aux circonstances physiques de la couverture de neige dans la nature. Outre les points de vue hydrologiques aussi la conductibilité de la couverture de neige dans différentes conditions sera éclaircie en détail; ce fait serait d'une grande importance en ce qui concerne les conditions de température du sol et celles de glace des systèmes d'eau.

1)
**REPORT OF THE RESEARCHES
CONCERNING SNOW- AND ICECONDITIONS,
WHICH HAVE BEEN MADE IN FINLAND AFTER 1936**

by

V. V. KORHONEN (Finlande)

To the meeting in Edinburg in 1936 I made a report of the researches concerning the snow- and ice conditions, which have been made in Finland up to the year 1936. Next performance is a continuation of it. As previously, the popular references and the yearbooks are omitted.

R. JURVA. *Tidsanalysen av isvintrarnas förlopp i haven kring Finland.* (Time analysis of the course of the ice-winters in the seas surrounding Finland). XIX Meeting of Scandinavian naturalists at Helsinki in 1936. Helsinki 1936.

R. JURVA. *Meriemme jäätälven pituuden ja jäätyminen lajuuden vaihteluista normaalitalveen verrattuna.* (The variations of the length of the ice-winter and the extent of the freezing of our seas when compared with the normal winter.) Terra 49:1. Helsinki 1937.

R. JURVA. *Laskelmia meriemme lämpövarastosta.* Calculations on the heat stock of our seas). Suomi merellä No 12. Helsinki 1938.

R. JURVA. *Laajimman jäätyminen vaihteluista ja jäätälven 1939/40 kulusta merissämme.* (Variations of the extensive freezing and the course of the winter 1939/40 in our seas). Laivastolehti N:o 6. Helsinki 1940.

R. JURVA. *Eräs kirjaanpano 1500-luvun keskipaikkeilta Helsingin ulkopuolella olevan meren jäätymisestä.* (A memorandum from the middle of the 16th century of the freezing of the sea outside Helsinki). Laivastolehti N:o 2. Helsinki 1941.

R. JURVA. *Vaikeista jäätälvista ja talven 1939/40 vaiheista.* (The difficult ice-winters and the stages of the winter 1939/40). Suomen Laivaston vuosikirja N:o IV. Helsinki 1941.

R. JURVA. *Suomenlahden merenjään liikunnasta sydäntälven aikana.* (The movement of the sea-ice of the Gulf of Finland during the depth of winter). Laivastolehti N:o 2. Helsinki 1942.

R. JURVA. *Itämeren ja sen lahtien laajimman jäätyminen ja auringonpilkkujen esiintymisen vaihteluista.* (Variations of the greatest freezing of the Baltic and its bays as well as those of the sunspots). Terra N:o 54. Helsinki 1942.

R. JURVA. *Über den allgemeinen Verlauf des Eiswinters in den Meeren Finnlands und über die Schwankungen der grössten Vereisung.* Sitzungsberichte d. Finnischen Akademie d. Wissenschaften 1941. Helsinki 1944.

R. JURVA. *Merén viimesyksyisestä jäätymisestä ja talven vaiheista huhtikuun alkuun.* (On the cooling of the sea during last autumn and the phases of the winter to the beginning of April). Laivastolehti 1945. Helsinki 1945.

R. JURVA. *Baltianmerestä, sen vesirungon rakenteesta, lämpöoliosista ja toiminnasta suurena lämmönlähteenä ympäristölleen.* (On the Baltic, the structure of its water body, its temperature conditions and signification as a heat source for the surroundings). Laivastolehti 1945. Helsinki 1945.

R. JURVA. *Ueber die grösste Geschwindigkeit der Vereisung des Meeres.* Sitzungsberichte d. Finnischen Akademie d. Wissenschaften 1946. Helsinki.

G. GRANQVIST. *Jäänlaadut.* — De olika isslagen. Merentutkimuslaitos — Havsforskningsinstitutet. Helsinki 1937.

G. GRANQVIST, R. JURVA. *Suomen jäätyö — Istjänsten i Finland.* M.t.I. — H.i., Helsinki 1937.

G. GRANQVIST. *Den baltiska isveckan 12-18 febr.1938.* Terra 50:4. Helsinki 1938.

P. KAITERA. *Lumen kevätsulamisesta ja sen vaikutuksesta vesiväylien purkautumissuhteisiin Suomessa.* (Spring melting of snow and its effect on the discharging ratios of watercourses in Finland. Thesis at the Finnish Technical High School at Helsinki 1939).

V. V. KORHONEN and T. O. LAVILA. *Die Periode mit Schneefällen in Finlaad.* Ann. Acad. Scient. Fennia Ser. A. Tom LI. No 2, Helsinki 1939.

V. V. KORHONEN. *Die Verteilung der Niederschläge, besonders der Schneefälle auf die verschiedenen Windrichtungen in Finnland.* Ann. Ac. Se. Fennicae. Ser.A.I.N. No 13. Helsinki 1942.

V. V. KORHONEN. *Ein neues Verfahren bei der Korrektion der Schneemessungen.* Ann. Ac. Se. Fennicae. Ser. A I. No 24. Helsinki 1944.

H. SIMOJOKI. *Ueber die Eisverhältnisse der Binnenseen Finnlands.* Ann. Ac. Se. Fennicae, Ser. A. Tom. LII, No 6. Helsinki 1940.

H. SIMOJOKI. *Über die Temperaturverhältnisse der Finnischen Seen im Winter.* Fennia 67. Helsinki 1940.

H. SIMOJOKI. *Über den Zeitpunkt des Entstehens und des Verschwindens der dauernder Schneedecke in Finnland.* Fennia 70, No 3, Helsinki 1947.

M. VÄRE. *Maan vesisuhteista ja viljelyskasvien sadoista Maasojan vesitaloudellisella koekentällä vuosina 1939-1944.* (Über die Wasserverhältnisse des Bodens und die Erträge von Kulturpflanzen auf dem wasserwirtschaftlichen Versuchsfeld Maasoja in den Jahren 1939-1944. Maa — ja vesiteknillisiä tutkimuksia No 5, Helsinki 1947.

2) SUR LES CONDITIONS DE GLACE DE LA BALTIQUE LE LONG DES COTES DE FINLANDE

par
R. JURVA (Finlande)

Ueber die Eisverhältnisse des Baltischen Meeres an den Küsten Finnlands.
Merentutkimuslaitoksen julkaisu — Havsforskningsinstitutets skrift N:o 114; Fennia 64, N:o 1; Mitteilung 13A der hydrologischen Konferenz der Baltischen Staaten
1936. 248 p., 95 maps and figures. Helsinki 1937.

This volume comprises, together with the map work « Atlas der Eisverhältnisse des Baltischen Meeres an den Küsten Finnlands » (see the following report), the two first parts of a planned extensive explanation of the ice conditions of the seas bordering on Finland.

Besides general comments the first part of the volume contains an explanation of the material used; its quality and composition are presented chronologically, following the historical development of the ice observation work. Thereafter follows a presentation of the material used, i.e. the Friday ice situations, by means of which the actual explanation of the ice conditions has been executed.

The second part of the volume forms the most important part of the investigation. In the beginning therein is shown how by using as such the Friday ice situation maps of the Institute for Marine Research it is possible to determine the general course of the ice winter at the coasts of Finland and in the seas bordering on Finland. For this purpose one had, however, to first develop an own, special cartographic method, based on the big structure of the ice situations, in which the elements characterizing the ice situations would be all considered. Only in this way it was possible to build such a normal winter, that depicted the general course of an ice winter, which could present all the fluctuations of the winters as well in regard to time as in respect to the extent of the freezing and at the same time even gives in average pattern a correct idea of the general character of the ice conditions of the different stages of the winter. As by means of the normal winter one another and as by means of it the general analysis of the winters would have to be performed, the **normal stages**, showing the different parts of a normal winter should have a « mobility » corresponding as precisely as possible to the actual conditions.

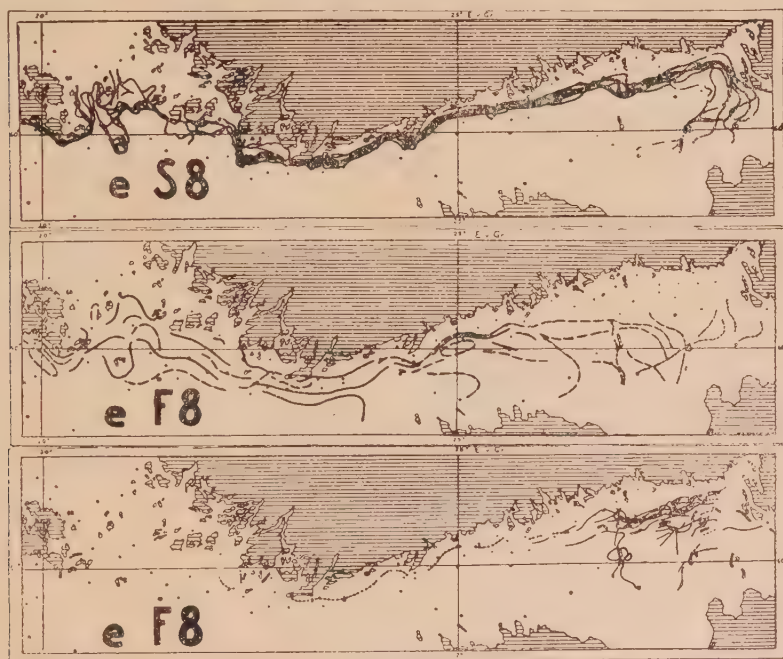


FIG. 1.

The normal stages obtain these characteristics in such a way that they are as far as possible made up of similar ice conditions quite independent of the time during which the situation had appeared in different winters — however with the limitation that each time on the one hand only the situations of the growing period of the ice and on the other those of the period of ice's diminishment were regarded as similar ones. By grouping the ice situations in this manner (Fig. 1) one then make up by means of the limits of the fast ice along the coast, of the edges of the loose ice and of the icefree areas as well as by means of the thickness of the ice, etc. the normal or S-stages (Fig. 2) of the different seas, which together give the general course of the normal winter. When the S-stages of all the seas had been determined the time was once more coupled to the event. This happened in such a manner that the appearance time of each stage corresponded to the average of the time interval in which the ice situations in question had appeared. From this time interval one determined too the stages of the very early, early, medium, retarded and very retarded appearance points.



FIG. 2

In each S-stage the determined curves show the edges of the fast ice along the coast; other curves again, the so-called frequency curves, represent the loose and frozen together ice outside the edge of the fast ice.

The number of the S-stages in the different seas are as follows : The Bothnian Bay 26, the Qvark 24, the Bothnian Sea 26, the Archipelago, the Aaland Sea and the North Baltic altogether 25 and the Gulf of Finland 28.

The new step was to made with the help of the S-stages the general or G-stages (Fig. 3) of the whole sea area. There were altogether 30 of these and they represent as such the medium long and medium difficult winter, i.e. a winter the annual appearance probability of which is $1/2$.



FIG. 3



FIG. 4

From the G-stages are further determined their appearance probability and, by means of the edges of the fast ice border and the frequency curves one could determine the time when the ice has been at some point and thereafter a number of « normal » winters of different character and length.

Of the many presented results of the methods of the investigation it may be mentioned only the graphic method developed from the time diagram, by means of which the time analysis of the general course of the ice winter is to be performed, also the presentation of the results gained thereby, and the every winter probabilities of the appearance of the different ice situations. They are, rounded off, for the exceptionnally early winter 1/20, very early 1/10, early 1/5, medium 1/3, retarded 1/5, very retarded 1/10 and exceptionnally retarded 1/20.



FIG 5.

Atlas der Eisverhältnisse des Baltischen Meeres an den Küsten Finnlands. Supplement to the publications Merentutkimuslaitoksen julkaisu — Havsforsknings-institutets skrift N:o 114; Fennia 64, N:o 1; and Mitt. 13A d. Hydrol. Konf. d. Balt. Staaten. III + 51 pages, folio. Helsinki 1937.

The Atlas, that is in direct connection to the above referred work « Ueber die Eisverhältnisse des Baltischen Meeres an den Küsten Finnlands » contains the normal stages of the general course of the ice winter, the ice thickness curves and the maps « Zeit mit Eis » (Time with ice) (Fig. 4). At first there are presented the normal or S-stages (Fig. 5) of the different seas, thereafter the general of G-stages comprising the whole of the sea area. Of the maps « Zeit mit Eis » the first two present the length in weeks of a medium long winter and the variations of the ice week figures. The following maps show the ice week figures at different ice probability during a very short, medium long and very long winter. In the foreword of the Atlas is given a closer account as to how the S-stages are made up — and of them the G-stages (Fig. 6) — and also how the curves of the maps are led, that present the ice week figures. Further we find in what a manner the appearance times of the stages and the annual appearance probability are determined and how the ice thickness figures and curves are obtained; what kind of a structure the border of the fast ice as well as how the so called frequency figures, and the frequency curves drawn by their help, are calculated, also what the used ice signs denote.

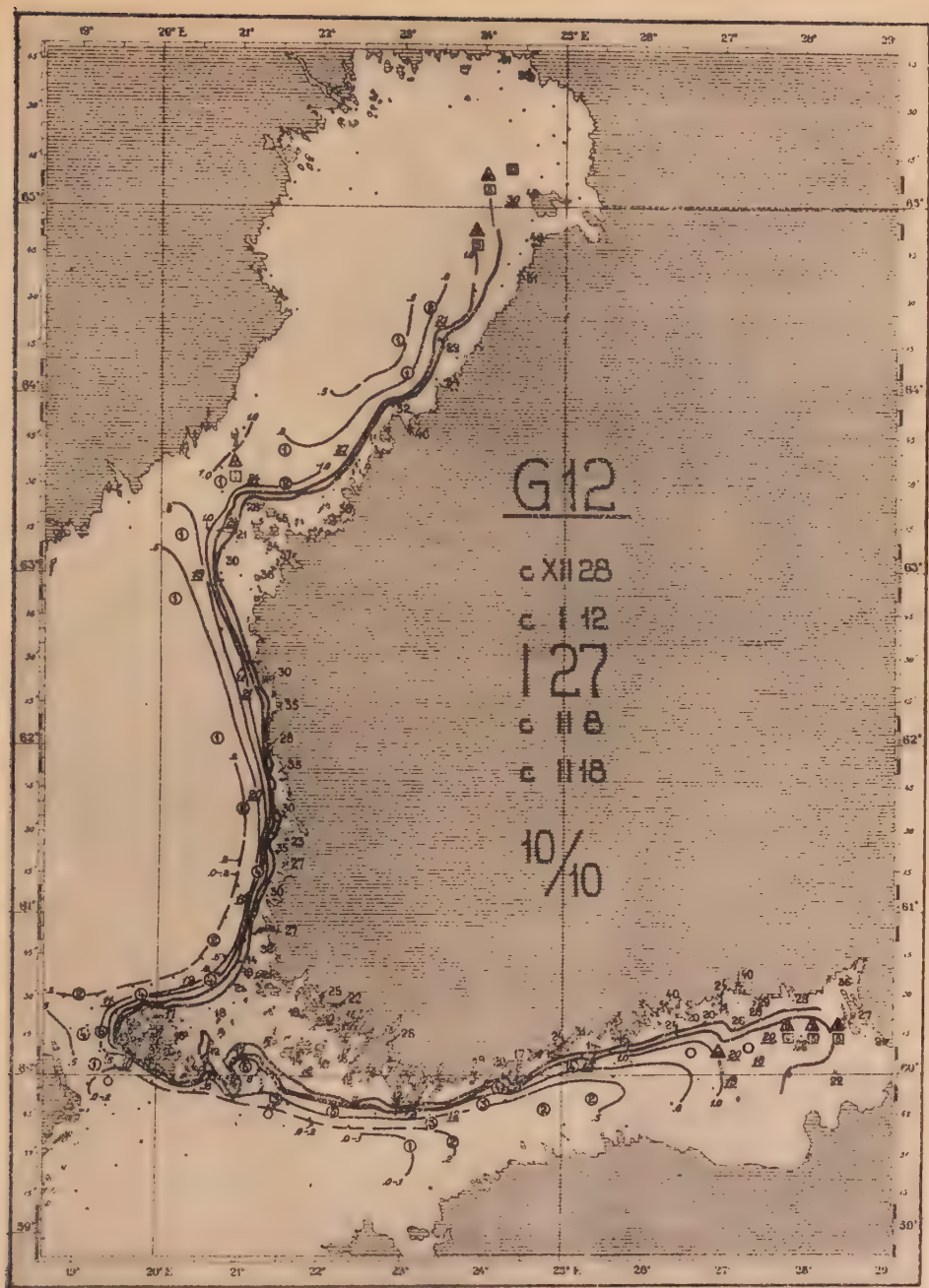


FIG. 6

3) SNOW MELTING IN THE SPRING AND ITS EFFECTS ON THE OPENING OF THE WATERWAYS IN FINLAND

by

Pentti KAITERA (Finlande)

Investigations were conducted in the period 1934-1937 during the Spring and beginning on the date of March 10th, every fifth day on observation stretches the length of which were 150 meters, and situated in different topographical sites in the whole country. Along all these stretches the snow depth has been taken in 30 places situated 5 meters from each other. Moreover tests have been made to ascertain the density of snow in certain points along these lines, the points being five, and by using as an instrument the snow scales invented by Korhonen. The snow-depth has thus been taken on points situated on about 110 test-stretches and the snow density ascertained upon 25-30 such stretches. Also these tests have been continued and at present a survey of the results obtained during a period of 15 years is being completed.

During the experiments combinations have been effected in order to ascertain how much the snow volume varies in different woods and woodcovered areas. On small wooden plains the snow amount is most abundant or 35-40 % more than upon tilled grounds. In a dense pine-forest the snow-amount is smaller than the same upon tilled fields. But on the other hand — the snow volume is greater in open pine-woods and in woods of mixed trees, and especially so in pure hardwood forests, than upon the open plains and fields. In the spring the snow melts in the course of 5 days to 45/50 % of its maximum water value, in the course of 10 days to abt. 70 %, in 15 days to abt. 85 % and in 30 days to abt. 95 %, and this will occur when the melting effect reaches its maximum in the spring. Notwithstanding that the absolute values (stated in mm) are varying considerably owing to the snow volume, the approximate values in different parts of the country under different topographical conditions will be rather close to the above stated figures. The melting maximum will occur some days later in the woods than on the fields, and similar variations will also be ascertained depending on the fact whether a slope is facing south, north west or east. From measuring of the vaporization occurring on the surface of the snow cover, it was discovered that the temperature being -2.90°C the vaporized volume a day (e) was 0.41 mm, temperature being $+1.5^{\circ}\text{C}$, $e = 0.29$ mm, and temperature being $+4.3^{\circ}\text{C}$, $e = 0.07$ mm. The reduction of the vaporization under temperature increases depends on the fact that during warmer periods there is likely to occur such days frequently when the air dew-point is higher than the temperature of the snow cover, thus causing aqueous vapours to tighten upon the snow cover.

4) CORRECTION DES MESURAGES DES CHUTES DE NEIGE FAITE SUR LA BASE DE LA NATURE DE LA PLACE DE MESURAGE

par

V. V. KORHONEN (Finlande)

Il est reconnu que le pluviomètre ne mesure pas exactement la quantité de pluie tombée. Pour cette raison on a étudié l'erreur faite dans les mesures et on a essayé de la préciser quantitativement. Pour ce qui concerne les pluies cela doit être fait par les mesures proportionnelles en plaçant près d'un pluviomètre installé d'une manière ordinaire, un pluviomètre dont la surface de rassemblement est au même niveau que la terre et tel que le rejaillissement de l'eau de pluie dans celui-ci a été empêché par des moyens appropriés. La question est cependant tout à fait autre quand il s'agit des chutes de neige. Outre que les erreurs sont alors beaucoup plus grandes que quand il s'agit de la pluie, les comparaisons ne peuvent être faites de telle façon qu'un *autre* mesureur soit placé dans la terre avec sa surface de rassemblement au niveau de la couverture de neige, parce que le vent apporte toujours tant de neige légère, qu'il reste considérablement de la neige sur ce mesureur. Pour faire des comparaisons, il n'existe pas d'autre moyen que de placer une plaque légère sur la surface de la couverture de neige près du mesureur de manière qu'elle soit exactement au même niveau que la surface de la neige et en même temps que le mesureur est échangé, de mesurer la quantité de neige rassemblée sur la plaque ou d'observer par courts intervalles la teneur d'eau de la couverture de neige et de comparer la quantité de neige prise du pluviomètre avec l'augmentation de la teneur d'eau de la couverture de neige pendant le même temps. Quand on observe attentivement la nature du couvert du voisinage de la place de mesurage, et son influence sur les différentes directions du vent, une correction peut être faite de la façon suivante :

Parce que les chutes de neige ne sont pas égales par rapport à chaque point cardinal, on avait reparté de longues séries de chutes de neige suivant les directions du vent. Cela a été fait dans 20 stations environ, situées dans différentes parties de la Finlande et les valeurs obtenues ont été rassemblées sur des cartes, une carte ayant été attribuée à chaque point cardinal. Sur la base de ces fiches, il est facile d'interpoler n'importe quelle place dans la répartition des chutes de neige par points cardinaux. (en pourcentage de la quantité totale)

Le voisinage de la position de la place a été ensuite calculé par les chiffres suivants qui dans le calcul définitif se présentent comme coefficients :

voisinage	coefficient
Forêt (ou endroit tout à fait protégé)	0
Enclos, buissons ou arbres	1/4
Bâtiments et derrière des endroits découverts	1/2
Endroit découvert et pas protégé	1

Ensuite on a multiplié la quantité relative des chutes de neige par les coefficients correspondants et les résultats ont été additionnés selon l'exemple suivant :

point cardinal	coefficient	quantité proportionnelle de neige	correction
N	0	9 %	0
NE	0	11	0
E	1	17	17
SE	1/2	18	9
S	1/2	14	7
SW	1/2	10	5
W	0	11	0
NW	0	10	0
		Total	38

Comme la répartition entre tous les points cardinaux a été donnée directement dans les pourcentages de toute quantité de neige, la somme donne le pourcentage dont le résultat de mesurage de la place doit être corrigé. Les comparaisons montrent, qu'on doit encore ajouter à cette somme le chiffre 3. Dans la station en question les chutes de neige doivent être corrigées de 41 %.

Afin de pouvoir corriger la quantité de pluie de tout un mois ou un an, il est nécessaire de connaître encore la part des chutes de neige dans la quantité totale de pluie. Si cela est connu, la correction peut être faite de telle façon que la méthode ci-dessus soit appliquée à la partie de pluie tombée comme neige.

Ce calcul de correction est simple et il a donné de bons résultats dans plusieurs recherches. Il doit être développé encore, en premier lieu en ce sens que les descriptions de position de la place doivent être plus détaillées et aussi en considérant le voisinage plus lointain de la place de mesurage : le coefficient de correction qui ne comporte ici qu'une échelle de 4 degrés doit être considéré seulement comme une première évaluation grossière et approximative, et il sera précisé par de nouvelles recherches basées sur une documentation encore plus étendue.

5) FLOODING DUE TO THE MELTING OF SNOW AND ICE

by

Capt. W. N. Mc CLEAN, M.A., A.M. Inst. C.E.

There is little reliable information as to snowfall on areas, and no good snow gauge appears to be available.

15 years of record on the Aberdeenshire Dee show, during the months of March and April, an average excess of 1.0 inch of Run-off over a Rainfall of 5.4 inches in the two months. On the rivers of the Ness and Lochy basins, February, March and April together, give an average excess of about 0.3 inch on a total rainfall of 16 to 20 inches. On the Dee, in March 1946, 3 days with maximum temperatures of 52° gave a total Run-off of 1.39 inches, with a peak at the rate of about 0.8 inch per day; and the accompanying rainfall may have totalled 0.2 inch in the 3 days.

The cold spell from January to March of 1947 resulted in serious floods on the Great Ouse, on the Trent and on the Lee. Estimates of the snow cover have been made; and the melting of this surface storage could not be absorbed by the storage areas available and resulted in serious flooding. As yet, there is no actual continuous record of Rainfall, Run-off and storage during this period; and on these large areas the measurement of Rainfall or Snowfall is no easy matter.

Good figures should be available for many water supply areas, where there is ample storage in Reservoirs. Also, small natural lakes, near the source of rivers, should provide good figures, with the necessary equipment and record.

6) PHYSICAL CHANGES IN SNOW-COVER LEADING TO RUNOFF, ESPECIALLY TO FLOODS

by

R. W. GERDEL (United States)

ABSTRACT

Technical papers and some unpublished data from a number of snow research laboratories established in the United States and from numerous other sources have been reviewed in this report.

The lack of a uniform definition and interpretation of technical terms applicable to snow research makes it difficult to resolve the opinions and conclusions of many authors into a factual analysis of the affect of the physical properties of the snowpack on runoff from melting snow or storm rainfall on the snowpack.

Most authorities agree that appreciable discharge of melt water from a snowpack does not occur until the pack becomes ripe. However, the term, ripe, does not appear to define a specific, or readily measurable condition of the pack. Various investigators have reported densities of from 25 % to 55 % for ripe snowpacks from which melt water was being discharged. Some reports discuss the homogeneous structure of a ripe pack while others indicate that there is little or no homogeneity present in a ripe pack at the peak of the melting season.

The liquid water storage capacity of a ripe snow of 40 % density has been reported to be as little as 3 % and as great as 50 % by weight of snow. Water transmission rates for ripe snow are reported as varying from 0.2 to 4.6 inches of snow per minute.

There appears to be a general opinion that new fallen, low density snow will effectively retard runoff from storm rainfall. However, reported experiments indicate that water will move through eleven inches of 10 % density snow in one minute while at 50 % density free water will move through only 4.6 inches of snow in one minute.

The conversion of the stellar, plate-like, or powder crystals associated with new fallen snow to the asymmetrical, single grain crystal as a part of the ripening process appears to be universally accepted. It is possible that such metamorphosis can be attributed, in part, to the mechanical energy derived from the tension of surface films of water on the snow crystals. Reduction in the surface area of the snow crystals appears to be associated with a reorientation of the resulting asymmetrical crystals with a reduction in surface tension capacity, an increase in capillary capacity and a reduction in non-capillary voids in the ripe snowpack. The combined surface tension and capillary capacity of a ripe snowpack appears to be very low. Such a pack which contained an unfrozen water content of 12.8 % during the peak melting period of the day was found to contain only 0.7 % of free water the following morning under conditions where little or no freezing occurred during the night.

It is hoped that the several snow research laboratories which have been recently established in the western United States may solve some of the problems associated with the physical properties of melting snow and the relation to runoff from snow melt and storm rainfall on snow.

Introduction

It is gratifying to report that during the past eight years several research projects were organized and implemented in the United States, particularly in the far western states, for the specific study of the physical and thermodynamic processes which are associated with normal and excessive rates of snow melt. Several of these projects are designed to obtain data which it is hoped will lead to improvements in the forecasting of runoff resulting from storm rainfall on snow cover as well as from snow melt.

During the past several years reports and technical papers from the following snow research projects have been published :

1. Crater Lake, Oregon, Snow Project. Soil Conservation Service, R. A. Work, Supervisor, Medford, Oregon.

2. Michigan Hydrologic Research Project. Michigan State College of Agriculture and Soil Conservation Service Cooperating. East Lansing, Michigan.

3. Intermountain Forest and Range Experiment Station, A. R. Croft, Conservationist, United States Forest Service, Ogden, Utah.

4. Rocky Mountain Forest and Range Experiment Station, Fraser Research Laboratory. H. G. Wilm, Silviculturist. United States Forest Services, Fort Collins, Colorado.

5. Soda Springs Cooperative Snow Research Project. University of Nevada and United States Weather Bureau Cooperating. Dr. J. E. Church, Meteorologist, University of Nevada, Reno, Nevada.

6. Cooperative Snow investigations. A series of snow research laboratories each consisting of a well instrumented basin and laboratory for physical research at high altitudes in the Rocky Mountains of Montana, the Cascade Mountains of Oregon and the Central Sierra Nevada of California.

These laboratories are operated jointly by the Corps of Engineers, Department of the Army and the United States Weather Bureau with limited cooperation by the United States Bureau of Reclamation, the United States Geological Survey and the United States Forest Service. Administrative and Analytical Offices are maintained at the Oakland Army Base, Oakland, California.

Many of the publications from these snow research projects have been reviewed for this report and are included in the bibliography.

Physical Changes in the Snowpack

Most authorities agree that before appreciable quantities of water may be discharged from a snowpack both the individual crystals and the entire crystal matrix of the pack undergo a distinct metamorphosis. This appears to be true regardless of the source of water, whether from melting snow or from rain upon the pack.

New fallen snow is usually composed of stellar, hexagonal platelike or light powdery crystals which form a matrix having a density of less than twenty percent and frequently less than seven percent.

The literature on snow contains little information upon the process by which these stellar, plate-like or powdery snow crystals are converted to the single grain asymmetrical crystals characteristic of a high density discharging snowpack.

The term « ripe », which has been applied by many writers to the characteristic crystal habit, matrix structure, isothermal character and high density of a typical melting spring snowpack, will be used in this report.

A ripe snowpack may be defined as one in which the balance between the morphological characteristics and thermodynamic potentials of the pack is so critical that a unit increase in available heat will result in a unit reduction in the water equivalent of the snowpack. This does not imply a unit degree day factor since under field conditions the source of heat may be other than that measured with an air thermometer.

The physical characteristics which are associated with a ripe snowpack appear to be the subject of considerable difference in opinion amongst investigators. Church (1) refers to the tendency of ripening snow to become homogeneous in density, however, he reports variations in density between 35.4 and 56.5 percent for melting snow. Clyde (3) found variations between 25 and 49 percent for melting snowpacks. Both of these investigators report a decrease in density after the start of meltwater discharge from the snowpack. Church (1) attributes this decrease in density to a reduction in capillary capacity.

Work and Frost (4) have shown that under the conditions at Crater Lake, Oregon, with a maximum snowpack of 130 inches, the weight of the overlying layers of snow had little or no effect in producing an increased density in the lower layers. They report that at the start of active melting of the snowpack the density varied in different layers between 41 and 55 percent and that there was a constant increase in density of all layers in the pack throughout the melting season.

However, homogeneous density was never attained since near the end of the season the density of different layers in the snowpack varied between 50 and 56 percent. These authors believe that atmospheric conditions at the time that the snow

is deposited may have an important bearing upon the ultimate density achieved by the horizon produced from such a deposit.

Observations at the Central Sierra Snow Laboratory near Donner Summit, California, one of the several laboratories operated by the Corps of Engineers and the U. S. Weather Bureau, indicate that even in a single storm several snow layers of different density and different crystal habit and profile structure may be deposited. Such layers may be sufficiently persistent to be identified months later in a ripe pack. The presence of many layers in an old snowpack remnant is shown in *Figure 2*. In this photograph, from the Central Sierra Snow Laboratory, the dark areas consist of fuchsine dyed melt water which is moving through a ripe snowpack from the dye salt on the snow surface to the right of the gage stick. The phenomena of melt water movement through the pack will be discussed in more detail later in this report.

From published reports, it appears that the integrated density of ripe snowpack may vary from 25 % to more than 55 %. Some of this variation may be due to differences in the time and method of sampling. Gerdel (5), (8) has shown that a melting ripe snowpack may have an unfrozen water content of 12.8 % by weight at 5.00 p.m. in the evening and by 9.00 a.m. the following morning contain only 0.7 % of free water. Drainage during the night apparently accounted for the entire loss of water from the pack since time freezing was confined to the surface inch of snow.

Such data indicate that a snowpack of 50 % density at 5.00 p.m. might, by the following morning, be reduced to a density of about 43 % as the result of drainage of free water from the pack during the night. The time of the day when snow surveys are made may have a considerable effect upon the reported density. However, differences in the density of ripe snowpacks cannot be attributed entirely to variations in the unfrozen water content of the pack at the time of sampling since the maximum unfrozen water content of a ripe melting snowpack in the western United States does not appear to exceed about 15 % by weight of snow (5, 6).

Church (1) presents data which indicate that the density of a ripe snowpack may be associated with elevation. He reports a ripe density of 42.2 % for an elevation of 6200 feet, 49.6 % at 8000 feet and 53.4 % at 9000 to 10,800 feet. However, he attributes such variation in density to the effect of wind-packing at the higher elevations. His data indicate also that at a single station the density of a ripe snowpack may vary from less than 40 % in one year to more than 50 % in another year.

Data obtained at the Central Sierra Snow Laboratory (14) show that during the late winter and spring of 1946 there was a consistent increase in snow density from 41 percent on April 3 to 54 percent on May 1, but that in mid-May at the height of the runoff season there was a range of 10 percent in average density between sample sites in the laboratory basin. The report from this laboratory states that "... although density has come to be regarded as a conservative element it is far from uniform."

The writer has reviewed the records of the California Cooperative Snow Surveys for ten or more years and has found no indication of a probable correlation between ripe pack density and elevation. On the other hand, the data do indicate a marked difference in ripe pack density from year to year.

Heat, the primary source of energy required for the ripening of snow is largely derived from turbulence exchange from the air and from solar radiation. Percolating melt water probably supplies the major energy for the ripening process at considerable depths in the snowpack.

Computations based upon equations and data presented by Sokolovsky (18), Wilson (13), and Gerdel (7) indicate that under the most favorable conditions only 0.20 to 0.34 inches of melt water per hour will be produced by heat available from turbulence exchange and solar radiation. The high water transmission rates of snow, which will be discussed in more detail later, preclude the retainment of any appreciable amounts of melt water on the snow surface where additional heat might be absorbed. Since the melt water is in constant contact with snow crystals at 32° F it is not possible to conceive of heat being transported by the small amount of melt water available to appreciable depths within the snowpack.

The conversion of low density, fluffy or stellar crystals to single grain, asymmetrical crystals common to a ripe pack may be attributed, partly, to the refreezing of water films on the surface of the crystals in horizons where temperatures may be below 32° F. A snowpack which is isothermal at 32° F during the day may lose sufficient heat at night as the result of outgoing long-wave radiation and diffusion to permit the freezing of water at considerable depths within the pack.

It has been frequently observed by the author of this report that the stellar, low density crystals of new fallen snow may be changed to the single grain, dense crystal under conditions where heat losses from the snow would not appear to be possible. In such cases it is probable that mechanical forces induced by the tension of the film water on the low density stellar crystals induces a shrinkage of the crystal to the maximum density, minimum volume and minimum surface area compatible with the crystal habits of ice. Where there is a high degree of contact between several crystals mechanical forces may produce one large high density crystal from a number of low density crystals.

It is probable that the mechanical forces induced by tension phenomena may take place in a low density snowpack to all depths to which surface melt water or rain water may percolate. It is probable also that when ambient temperatures are below freezing water films can be produced by radiant heat on snow crystals at considerable depths in the pack. Gerdel (7) has shown that as much as 10 gm cal/cm²/day of solar radiation may be transmitted to depths of 10 inches in the snow and that measurable amounts of radiant heat were found at depths of 16 inches in the snow on cloud-free days.

In order to determine the effect of the mechanical forces associated with capillary and surface water films on structural and density changes in the snowpack several experiments were performed at the Central Snow Laboratory.

The procedure used in these experiments consist of the removal of a core of snow from the pack by means of a sampler 8 inches in diameter and 25 inches in length which was provided with a removable screen cap at the lower end. The volume and density of the snow core were measured and two inches of water at 32° F added uniformly over the exposed upper end of the core. Samples of snow were taken adjacent to the site from which the core was removed and the unfrozen water content of the snow determined by the thermal quality procedure described by Bernard and Wilson (16). The core of snow still contained in the sampler was replaced in the snowpack and good contact established through the screen cap, between the bottom of the core and the snow from which it had been removed.

Newspapers and a foot or more of snow were used to cover the sample to prevent freezing or thawing during the subsequent three hour drainage period.

At the end of the drainage period the core and sampler were removed from the snowpack, the volume and density measured and samples removed from the core for the calorimetric determination of the unfrozen water content.

The experiments were conducted on days when air temperature were above freezing and the snowpack isothermal at 32° F.

Previously unpublished data from these experiments which are presented in Table 1 are not considered sufficient to justify definite conclusions. They do indicate that storm rainfall on low density snow may cause a rapid change in the morphological characteristics of the snowpack and that any detention of appreciable amounts of freewater must be of a very temporary nature. They indicate also that the metamorphosis of the snowpack is not dependent alone upon alternate thawing and freezing since all changes in density were induced by water at 32° F applied to snow maintained at 32° F.

Visual inspection of the low density (14.4 %) snow before the start of the experiment on March 4 indicated that the entire column was composed of a very loose matrix of stellar type crystals typical to new snow. When the core was examined at the end of the three-hour drainage period, a major portion of the snow core was composed of very fine, asymmetrical, single grain crystals. On March 11 changes in density after the application of water appeared to be the result of some increase in crystal size but largely to the shrinkage of the core of snow. This shrinkage may be attributed to the tension created by film and capillary water drawing the crystals closer together, reducing the size of the non-capillary voids and increasing interfacial contact.

The Water Holding Capacity of Snow

A snowpack, isothermal at 32° F may retain free water by surface tension and capillary tension. A new fallen snow of low density, consisting of a very loose matrix of stellar crystals will have a relatively large surface tension capacity and a low capillary capacity. In such a snowpack the majority of the voids are too large and the porosity too great to permit any appreciable retention of capillary water. A ripe snowpack of approximately 50 percent density, composed largely

of medium to coarse asymmetrical crystals with a relatively high degree of facial contact will have a greater capillary capacity but lesser surface tension capacity than the low density, stellar crystal snow.

Various investigators who have reported on studies of the water holding capacity of snow have failed to recognize the difference between surface tension and capillary tension phenomena. Horton (10) states « New light snow of density 0.10 to 0.15 has a small capillary capacity. » However, he proceeds to cite as a theoretical example the retention of 0.6 inches of rainwater by a six-inch snow blanket of 10 percent density. This example infers a water holding capacity of 50 percent by weight of wet snow. Church (1) on the other hand, discusses the greater capillarity of new snow as compared to a ripe snowpack. The statement by Horton indicates a recognition of the difference between surface tension and capillary tension phenomena while Church apparently combines both phenomena under the term « capillarity ».

The literature in snow contains many other instances of apparent conflict in opinion on moisture tension in snow, most of which may be attributed to lack of satisfactory definition of the terms used to express the phenomena.

In a soil, the capillary and surface tension potentials are relatively constant and may be combined into a single term in any general consideration of the moisture capacity of a soil. On the other hand, the moisture holding capacity of snow is greatly influenced by the metamorphism which accompanies the ripening of the snowpack.

Some studies are reported by Gerdel and Codd (6) on the water holding capacity of a new fallen snow and an older ripe snowpack during a period of heavy rain. They found that a new snow of 10 percent density had a temporary water holding capacity of between 12.8 and 19.3 percent while the old snow of 38 percent density had a water holding capacity of 0.4 to 5.9 percent where free drainage of water was favored by sloping ground beneath the pack. Where water was impounded in depressions beneath the pack, the liquid water content of the new snow was 30.8 percent. This high liquid water content is not considered to represent the true water holding capacity of snow, but rather the surface detention capacity of the underlying impermeable strata.

The high water detention capacity which many writers (1, 2, 10, 12, 17) have associated with low density snow may be considered as a temporary phenomena associated with the large surface area of new fallen snow crystals. It is possible, also, that many of the normal drainage channels on the soil surface are blocked by the new fallen snow. Water moving through the low density snow in excess of the infiltration capacity of the soil may be impounded until drainage ways between the soil surface and the snow blanket are established.

Water Transmission Through Snow

Oechsli (19) applied fuchsine dyed water to snow of different densities. By timing the movement of the dyed water he found that it was transmitted through 11 inches of snow of 10 % density and 4.6 inches of snow of 50 % density in one minute.

The lowest value obtained by Oechsli is almost double the rate of 2.28 inches of snow per minute reported by Horton (11).

Gerdel (5) found that when 0.5 inches of water was applied to the surface of a ripe snowpack, almost one hour was required for all of the applied water to percolate through one foot of snow. This, however, is more indicative of the detention storage time than the water transmission rate of a snowpack.

By means of a dielectric snow moisture meter developed by Gerdel it has been possible to measure the drainage time of a snowpack when water is applied to the surface of the snow. Data obtained at the Central Sierra Snow Laboratory with this instrument, which have not been previously published, are presented graphically in Figure 1.

Since the data are from a single experiment only tentative conclusions are possible. However, the shape of the curve in Figure 1 indicate that the water transmission rate of a snow is a function of the amount of liquid water available. Of the two inches of water applied to the surface of the snow, one inch had passed through six inches of snow in ten minutes. In forty minutes only 1.5. inches of the applied water had passed through six inches of snow. It is apparent that approximately two hours would be required for the transmission of the entire two inches of applied water through the six inches surface layer of snow.

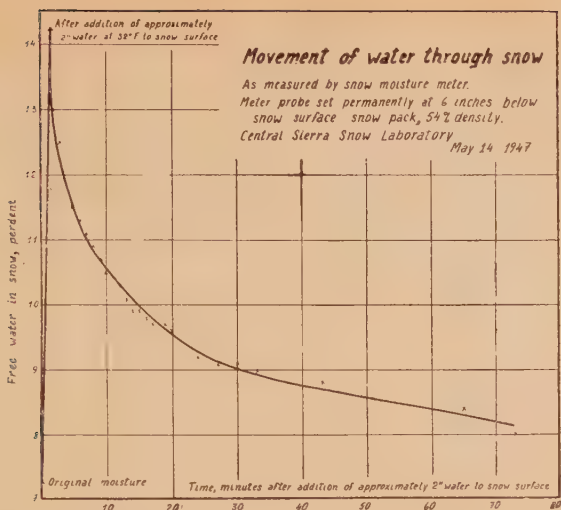


FIG. 1. Movement of Water Through Snow. As measured by the dielectric snow moisture meter at six inches below the surface of a 54 % density snowpack. Central Sierra Snow Laboratory May 14, 1947.

The flow pattern of liquid water through a ripe snowpack has been the subject of considerable investigation in the Central Sierra Laboratory. By means of the dye fuchsine, also known as rosanilin hydrochloride, the movement of melt or rain water can be traced through the snow. This dye, which is a dark purplish green in the absence of liquid water, changes to a brilliant red when free water is present in the snow. Since a visible tint is produced by as little as one part of the dye in ten million parts of water, it is possible to trace the movement of very small amounts of water through the snow.

This dye has the further advantage of not affecting the freezing point of the snow. In a saturated aqueous solution the depression of the freezing point is only 0.001° C.

Data from the project under investigation at the Central Sierra Snow Laboratory have not been published although a preliminary report has been prepared for the information of the cooperating agencies.

The dye in salt form was applied to the surface of a four-foot, midwinter snowpack at the start of a rain storm. The top half of this pack was new snow while the lower half was old, ripe snow. At the end of the storm which produced 2.14 inches of rain, the course of the water through the pack was determined by digging tranches down to the soil beneath the pack. It was found that the dyed water took a very devious course through both the new snow and the old, ripe snow. The point of contact of the dyed water with the soil was five feet, horizontally from the point of application of the dye in the snow surface.

During the spring melting season in the Sierra Nevada the dye in salt form was applied to the surface of an old, ripe snowpack in a wooded, well shaded area at a time when ambient temperatures were about 60° F. At the time of application of the dye the liquid water content of the snow was found by calorimetric procedures to be between 1.5 and 2.6 percent by weight. Variations in the liquid water content appeared to be associated with structural differences between definitely identifiable horizons in the pack.

About three hours after the application of the dye to the snow surface over an area about 6 inches by 10 inches, a pit was dug and a vertical face of the snow exposed. By use of filters and panchromatic film the movement of the dyed melt water through the snow was photographically recorded. The presence of melt water in the pack is indicated in the photographs by dark areas on the vertical face of the snow. The point of application of the dye is indicated also by a dark area on the snow surface. A rod, twenty-four inches in length and graduated in tenths of an inch was included in the photographs for reference purposes.

Table 1.—Effect of adding two inches of water at 32° F to a core of snow 8 inches in diameter and 25 inches in depth. Snow temperature was maintained at 32° F during the course of the experiment. Data show the physical properties of the snow *before* and 3 *hours after* the addition of water to the snow core.

Date	Core Volume Cu. Inch		Density Percent		Water Equivalent Inches		Increase		Unfrozen Water** Percent by Weight	
	Before	After	Before	After	Before	After	Percent	Before	Before	After
1947										
Mar.										
4	125	92	14.4	21.9	3.60	4.00	11.1	0.0	2.0	
11	125	90	26.4	37.9	6.55	6.83	4.3	0.0	0.0*	
22	125	125	39.0	40.4	9.76	10.10	3.5	7.4	5.5	

* Unfrozen water content determined calorimetrically.

** Snow appeared moist but quantity too small to measure with type of calorimeter used.

One of these photographs, Figure 2, vividly portrays the path of melt water through an old snowpack containing several semi-impervious layers. These layers commonly referred to as ice planes, appear to consist of coarse crystals cemented together to form a crust rather than a distinct plane of ice. Where the crust is impervious there is a temporary impounding of melt water. This phenomena is indicated by the darker lines just to the right of the scale rod at an elevation of sixteen inches above the soil surface. At least eight semi-permeable ice crusts can be identified in this photograph each of which appears to have retarded or diverted the vertical movement of the melt water.

A new vertical surface was exposed in the pit about four hours after the dye had been applied to the snow. The photograph taken at this time, Figure 3, shows that the most impervious planes have started to disintegrate and that an almost perpendicular transmission of melt-water appears to have been established. The presence of several vertical impervious planes in the snowpack also is indicated in this photograph.

Since the snowpack on which this experiment was undertaken represented a two-foot deep remnant of what had been a six-foot pack of approximately 50 percent density at the start of the melting season, it probably represented the highest degree of homogeneity, which may be expected in a ripe pack composed of the accumulated snow from many storms.

Observations at the Central Sierra Snow Laboratory (14) indicate that definite drainage may be established within a layer of new, low density snow during a period of rain. This condition is photographically illustrated in Figure 4.

The markings on the snow surface in this photograph were produced by 2.5 inches of rain on six feet of snow. The rain immediately followed a storm which had left several feet of new snow on top of an older but not ripe pack. Since no flow of water was observed on the surface of the snow during the rain, and the markings were not visible until several hours after the end of the rain, it may be concluded that flow channels were established within the snowpack and that when sufficient drainage had taken place the snow settled into these channels.

Drainage of Water From the Snowpack

In discussion the influence of mountain snowfields on stream flow in the Sacramento Valley, Fletscher (9) states, "The degree of density of the snow largely determines when substantial runoff due to melting will begin..." Garstka (15) refers to the possibility of coupling snow density data with certain meteorological observations as a means of forecasting the yield of water from snowfields.

The hydrometeorological log of the Central Sierra Snow Laboratory (14) shows that spring runoff from the laboratory basin did not begin until April 10, 1946 although the snowpack in the basin had reached an average density of 43 % by March 3. For the 1946 season an average density of 49 % appears to be associated with ripeness of the pack and beginning of runoff in this laboratory area.

Linsley (21) found that as the season advances there is a greater amount of runoff from the same amount of snow melt or the same number of degree days. He attributes this to a gradual ripeness of the snow.

Work and Frost (4) show also that there is a great increase in snow melt per degree day unit as the season progresses. They found that there was a variation between 0.027 inches of melt per degree day in March to 0.658 inches per degree day in early June. Their data do not indicate that changes in density of the pack are closely associated with the degree day factor. With an integrated pack density of 51.1 percent on April 28, they found a degree day factor of 0.123 for the succeeding two weeks. For the two-week period following May 26 when the integrated density of the pack was 52.7 percent, they report a degree day factor of 0.658. Present standard snow survey procedures would not produce reliable data within the 1.6 percent density differences between these two periods.

In view of the great variations in density of a ripe snowpack as reported by many investigators, it would appear to be very difficult to include snow density as a function in any runoff formula. This does not imply that there is no general relationship between density and the beginning of runoff. Snow with a density, of less than 30 % will usually produce no runoff while snow of more than 40 % may release an appreciable volume of water when ambient temperatures are above 32° or other sources of heat are sufficient to melt the snow.

Summary

Any effort to associate physical properties of the snowpack with runoff from melting snow or storm rainfall on the snowpack is complicated by the indefinite, often controversial statements contained in the literature on snow. Although the authorities agree that ripeness is a criterion of the beginning of important discharge of water from the snowpack, there appears to be no specific structural or density phenomena with ripeness other than crystal structure. The density of a ripe pack has been variously reported at from 25 to 55 %. Some refer to the homogeneous structure of a ripe pack and others have found no homogeneity present in a ripe pack at the peak of the melting season.

Practically all data on the water transmission rates in snow are based upon highly artificial conditions and the reported rates may be a function of the experimental technique rather than the physical condition of the snow upon which the experiments were performed. Water has been reported as moving through snow at rates varying between one foot in less than one minute to 1 foot in more than one hour.

For ripe snow 40 % density or greater the literature contains references to water transmission rates varying between 0.2 and 4.6 inches of snow per minute.

Many articles on the hydrology of snow covered watersheds indicate a general opinion that the presence of a new low density snow cover will effectively retard runoff from storm rainfall. Oechsli (19) shows, however, that water will be transmitted through 11 inches of new snow at 10 % density in one minute while at 50 % density water will move only through 4.6 inches of snow in one minute.

It is possible that the depressive effect of a new snow cover on runoff may be due largely to the temporary impounding of water on the soil beneath the snow mantle previous to the development of flow channels within the snow or between the snow and soil surface. If the soil surface is not frozen the impounding of water may result in an increased infiltration into the soil which also would depress the hydrograph resulting from rain on a snow covered basin.

The actual flowage of slush consisting of a considerable amount of snow suspended in or floating on water has been reported by Garstka (15) and Hüllinghorst (20). Both overland flow and bore flow in channels beneath the snow would be greatly retarded as a result of the increased viscosity of the slush. Such flowing slush within or beneath a snowpack may temporarily plug previously established channels and increase the temporary detention storage effect of a new snow mantle.

The writer has not been able to find a single reference in the literature in which runoff from storm rainfall on new, low density snow is compared with an equivalent rain on an old ripe pack equal in depth or equal in water equivalent to the new snow cover.

The literature in snow contains almost no references to long-time, intensive studies on the physical changes in the snowpack and the influence of such changes on runoff from melting or storm rainfall. The majority of the references which have been reviewed present opinions of individuals which were derived largely from visual observations and theoretical interpretations supplemented by a few very limited experiments. Very little of the experimental data which are available were obtained under natural conditions. The reported results of many such experiments may be related more to the technique used than to the physical properties of the snow under investigations.

It is not possible at this time to relate the physical changes in the snow cover to runoff of melt water or storm rainfall on the snow mantle. However, a number of the snow research projects and laboratories which have been recently established in the western United States should provide answers to such problems within the near future.

BIBLIOGRAPHY

1. CHURCH, J. E., *The Melting of Snow*. Proceedings Central Snow Conference, pp. 21-31, 1941.
2. CHURCH, J. E., *Report of the Committee on Snow*. Trans. Amer. Geophys. Union, p. 401, 1942.
3. CLYDE, George D., *Snow Melting Characteristics*. Tech. Bull. No. 231, Utah Agricultural Experiment Station, 1931.
4. WORK, R. A. and FORST, W. T., *Snow Layer Density Changes*. Western Snow Conference, April 21, 1947. Submitted for publication, Trans. Amer. Geophys. Union.
5. GERDEL, R. W., *Snow studies at the Cooperative Snow Research Project, Soda Springs, California*. Annual Report, 1943-1944.
6. GERDEL, R. W., and CODD, A. R., *Snow Studies at the Cooperative Snow Research Project, Soda Springs, California*. Annual Report, 1944-1945.
7. GERDEL, R. W., *Penetration of Radiation into the Snowpack*. Amer. Geophys. Union, North Pacific Section of Hydrology, Meetings April 22, 1947. Submitted for publication, Trans. Amer. Geophys. Union.
8. GERDEL, R. W., *The dynamics of Liquid Water in Deep Snowpacks*. Trans. Amer. Geophys. Union, pp. 83-90, 1945.
9. FLETCHER, Edgar H., *Melting Snow as a Flood Factor in the Sierra Nevada*. Bull. Amer. Meteorol. Soc. February, 1940.
10. HORTON, Robert E., *The Role of Snow, Ice and Frost in the Hydrologic Cycle*. Proceedings Central Snow Conference. pp. 5-21, 1941.
11. HORTON, Robert E., *The Melting of Snow*. Monthly Weather Review. 43 : 599-605, 1915.
12. WILSON, W. T., *Some Observations of the Thermal Quality of Snow*. Trans. Amer. Geophys. Union, pp. 553-556, 1942.
13. WILSON, W. T., *An Outline of the Thermodynamics of Snow Melt*. Trans. Amer. Geophys. Union, pp. 182-195, 1941.
14. Hydrometeorological Log of the Central Sierra Snow Laboratory, 1945-1946. *Cooperative Snow Investigations, Corps of Engineers and U. S. Weather Bureau*. Tech. Report No. 5, 1947.
15. GARSTKA, Walter U., *Hydrology of Small Watersheds Under Winter Conditions of Snow-Cover and Frozen Soil*. Trans. Amer. Geophys. Union, pp. 838-871, 1944.
16. BERNARD, Merrill and WILSON, Walter T., *A New Technique for the Determination of Heat necessary to Melt Snow*. Trans. Amer. Geophys. Union, pp. 178-181 1941.
17. PARSONS, Walter J., *Potential Floods in the Sacramento Valley*. Trans. Amer. Geophys. Union, pp. 106-110, 1941.
18. SOKOLOVSKY, D. L., *Effect of the Intensity of Snow Melt on the Creation of Maximum Discharges*. Trans. Internat. Comm. of Snow and Glaciers, 1939.
19. OECHSLIN, Max, *Wasserdurchsickerung and Wasserruckhaltvermogen der Schneedecke*. Trans. Internat. Comm. of Snow and Glaciers, 1939.
20. HULLINGHORST, David W., *Minutes of the Eastern Snow Conference*. Mimeo Report, 1941.
21. LINSLEY, Ray K., *A Simple Procedure for the Day-to-Day Forecasting of Run-off from Snow-Melt*. Trans. Amer. Geophys. Union. pp. 62-67, 1943.



FIG. 2. Melt Water Movement Through an Old Ripe Snowpack. Profile exposed about three hours after application of fuchsine dye to the snow surface to the right of the gage stick.
Central Sierra Snow Laboratory, May 15, 1947.

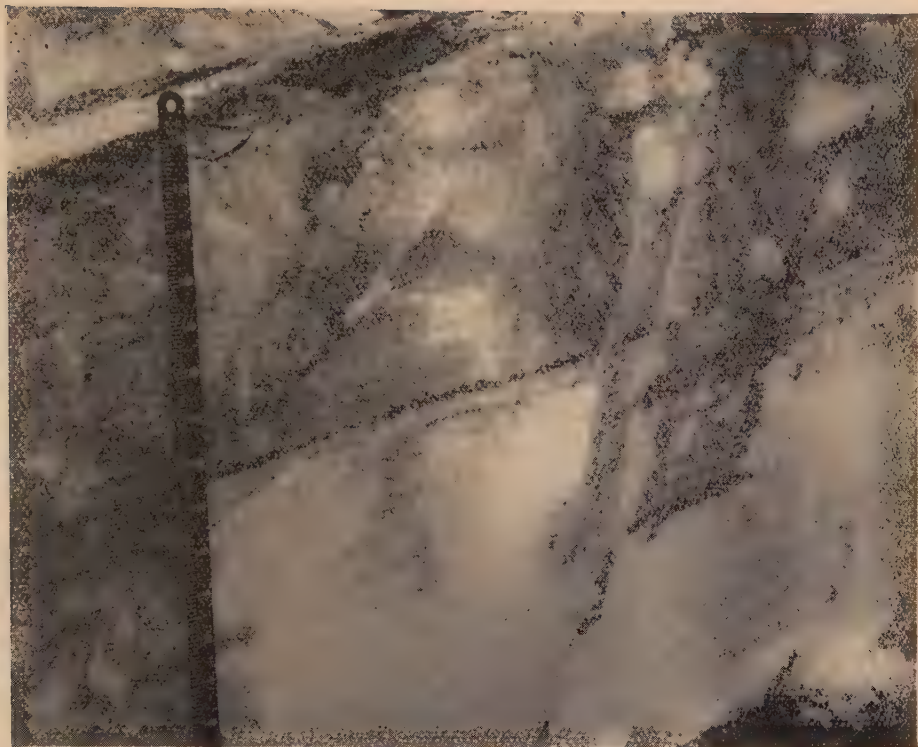


FIG. 3. Melt Water Movement Through an Old Ripe Snowpack. Profile exposed about four hours after addition of fuchsine dye to the snow surface.

Central Sierra Snow Laboratory, May 15, 1947.



FIG. 4. Effect of Heavy Rain on Snow. The channels and pits on the snow surface were not apparent until several hours after cessation of a heavy rain, they appear to be the result of subsidence into sub-surface channels through which water had flowed.

Central Sierra Snow Laboratory, December, 1946.

7) UEBER DEN ABBAU DER ALPINEN SCHNEEDECKE

von

Marcel de QUERVAIN

Eidg. Institut für Schnee- und Lawinenforschung Davos/Weissfluhjoch (Schweiz)

1. Einleitung

Am Eidg. Institut für Schnee- und Lawinenforschung sind in den letzten Jahren zahlreiche Messungen und Beobachtungen durchgeführt worden, die sich direkt oder indirekt auf den Abbau der Schneedecke beziehen. Es soll versucht werden, diese Einzelbeobachtungen unter Berücksichtigung allgemein bekannter Tatsachen in ihrem natürlichen Zusammenhang zu betrachten. Wir haben nicht die Absicht, hier umfassendes Zahlenmaterial zu veröffentlichen, vielmehr streben wir darnach, die *Gesetzmässigkeiten und Grössenordnungen* einzelner Erscheinungen an Hand von typischen Beispielen aufzuzeigen.

Dort, wo von Beobachtungen in der freien Natur die Rede ist, bezieht sich unser Referat auf die *alpine Schneedecke*, und zwar in erster Linie auf die Zone zwischen der Waldgrenze und der Firngrenze. In den Ostschweizeralpen verläuft die Waldgrenze in ca. 2000 m Höhe (6600 Ft), die Firngrenze in ca. 2800 m (9200 Ft). Das schweizerische Schneeforschungsinstitut liegt oberhalb Davos in 2660 m (8770 Ft) m. ü. M., also in der erwähnten Zone. Ganz allgemein sind in diesem Gebiet folgende Schneeeverhältnisse anzutreffen: Gewöhnlich setzen die ersten grösseren Schneefälle eines Winters, die zu einer geschlossenen Schneedecke führen, gegen Ende Oktober ein. Im März oder April werden die grössten Schneehöhen von 2-3 m gemessen. Gegen Ende April beginnt bereits der Abbau und anfangs Juli endlich verschwinden die letzten Schneereste im horizontalen, sonnenexponierten Versuchsfeld des Institutes. Die Bodenoberflächentemperaturen liegen im Hochwinter einige zehntel Grad, bis ein Grad (C°) unter dem Gefrierpunkt, während die Lufttemperaturen in derselben Zeit meistens zwischen -5° und -20° pendeln, gelegentlich unter -25° absinken, und selten einmal den Nullpunkt übersteigen.

2. Zum Wärmehaushalt der Schneedecke

Wenn man von sekundären Windverfrachtungen des Schnees absieht, steht der Abbau der Schneedecke ganz unter dem Regime seines *Wärmehaushaltes*. Dies betrifft sowohl die Schmelzwasserbildung, als auch die Verdunstung.

Der Wärmehaushalt wird bestimmt durch den Strahlungsaustausch und den direkten Wärmeübergang zwischen Schnee und Atmosphäre, durch die Wärmezufuhr vom Boden in die Schneedecke, ferner durch die Wärmetönung bei Verdunstung, Kondensation, Schmelzen und Gefrieren und schliesslich durch die Wärmeübertragung im Innern der Schneedecke. Eine Massenänderung soll in diesem Zusammenhang nicht berücksichtigt werden.

a) Strahlungsaustausch

Wir greifen auf einige Unterlagen zurück, die von O. Eckel und C. Thams [1] im Jahr 1939 veröffentlicht wurden, und die auf Messungen aus dem Gebiet von Davos beruhen.

Bei klarer Witterung erreicht die tägliche Globalstrahlung in den Monaten der beginnenden Schneeschmelze (März/April) Werte von 500 - 700 cal/cm² Tag. Davon wird ein Anteil von 0,5 bis 0,9, je nach dem Zustand der Schneeoberfläche, reflektiert.

Die für einen klaren Apriltag gemessene absorbierte Strahlung beträgt um 300 cal/cm² und vermag theoretisch rund gegen 40 mm Schmelzwasser zu erzeugen. Ob dieses Schmelzwasser auch wirklich entsteht und zum Abfluss gelangt, hängt u.a. vom augenblicklichen Zustand der Schneedecke ab.

Die Bedeutung der *Albedo* für den Wärmehaushalt und damit für den Abbau der Schneedecke ist erheblich. Dafür ein Beispiel: Im Frühling 1947 bedeckte eine

Schicht von gelbem Saharastaub die Schneeoberfläche und setzte die Albedo stark herab. In der Folge ging der Abbau der Schneedecke wesentlich schneller vor sich als in den vorangegangenen Jahren und war ungefähr 1 Monat früher beendet (Bild 1). Allerdings herrschten in dieser Zeit ungewöhnlich hohe Temperaturen und beschleunigten die Schneeschmelze. Trotzdem muss der überdurchschnittliche Schneeverlust zu einem guten Teil (schätzungsweise zu $1/3$ bis $1/2$ der Anomalie) der verminderten Albedo zu Last gelegt werden.

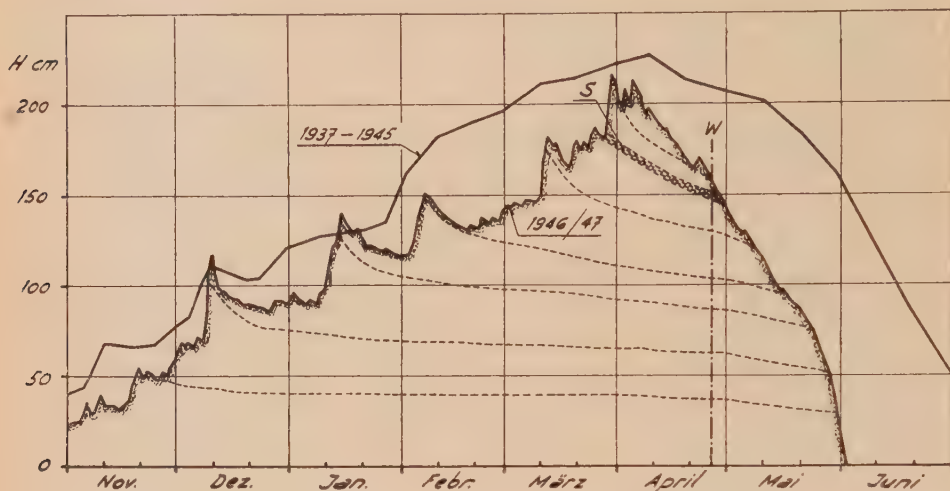


BILD 1. Schneehöhen im Winter 1946 und Durchschnittswerte der Winter 1937-1945
 S = Sahara-Staubschicht, gefallen am 29. März.
 W = Zeitpunkt des völligen Auftaus.
 Gestrichelte Linien: Setzungskurven einzelner Schichten (mit elektr. Pegel gemessen)

Die nicht reflektierte Strahlung dringt mit exponentiell abnehmender Intensität ins Innere der Schneedecke :

$$(1) \quad J(x) = J_0 e^{-kx}$$

Der Absorptionskoeffizient k liegt nach den Messungen von Eckel und Thams und den von ihnen zitierten Autoren im sichtbaren Spektrum zwischen $0,08$ und $0,2 \text{ cm}^{-1}$. Daraus folgt, dass in den obersten 10 cm der Schneedecke über 50% und bis in 50 cm Tiefe 99% der eindringenden Energie absorbiert werden. Beträgt also die Schneehöhe weniger als $40 - 50 \text{ cm}$, dringt ein merklicher Teil der Strahlung auf den Boden durch und heizt diesen auf. Durch die Rückheizung vom Boden her wird das Schmelzen im Schnee verstärkt. Daher hält die durch Strahlung bewirkte Schmelzwasser-erzeugung bis zum Verschwinden der Schneedecke fast unvermindert an.

Neben der Einstrahlung ist die nächtliche, langwellige *Ausstrahlung* der Schneeoberfläche für die Schneedeckenentwicklung wichtig. Sie bewirkt eine oberflächliche Unterkühlung, die in klaren Nächten nach unseren Beobachtungen 15°C unter die Lufttemperatur betragen kann. Den Bilanzmessungen von Eckel und Thams entnehmen wir, dass der nächtliche Ausstrahlungsverlust im April $80 - 90 \text{ cal/cm}^2$ erreichen kann. Dies ist ein guter Teil der am Tag aufgenommenen Einstrahlung von ca. 300 cal/cm^2 . Die Ausstrahlung übt also eine konservierende Wirkung auf die Schneedecke aus.

Kennzeichnend für das Verhalten der Schneeoberfläche unter der Ein- und Ausstrahlung im Frühling ist die Bildung der sog. Firnspiegel. Bei wiederholtem Aufschmelzen der Oberfläche entsteht eine glasklare, im Gegenlicht hell glänzende $1 - 2 \text{ mm}$ starke Eishaut. Durch das Zusammenwirken von Ausstrahlung und Verdunstungskälte wird diese Haut vor der Auflösung bewahrt. Die Einstrahlung dringt hindurch und beginnt die unmittelbar darunter liegende Schneeschicht aufzuweichen und abzubauen. Diese sinkt zusammen und löst sich dabei von der dünnen Eiskruste, dem Firnspiegel, stellenweise ab, sodass Hohlräume entstehen. Durch Kondensation von Dampf wird der Firnspiegel von unten ständig genährt und erneuert.



BILD 2. Firnspiegel am Rand einer Mulde.
Unterlage weggeschmolzen.

Bild 2 zeigt ein Stück eines Firnspiegels am Rande einer Mulde. Es ragt frei in die Luft, während die Unterlage bereits weggeschmolzen ist.

b) *Wärmeübergang zwischen Luft und Schnee.*

Ueber dieses Problem liegen neben den Untersudrungen von H.U. Sverdrup [11] nur einige wenige Anhaltspunkte vor, so indirekte Bestimmungen aus der gesamter Wärmebilanz von Angström [2] und von Niederdorfer [3]. Rechnerisch kann der Wärmeübergang bei einfachen, idealisierten Verhältnissen grössenordnungsmässig aus der Gleichung des Wärmeüberganges ermittelt werden.

$$(2) \quad W = \alpha (T_1 - T_2) \text{ kcal/m}^2 \cdot h$$

Diese setzt definierte Schneeoberflächen- und Lufttemperatur (T_2 und T_1) und einfache Luftströmungsverhältnisse voraus. Für die Wärmeübergangszahl α der Luft finden sich in der wärmetechnischen Literatur verschiedene, den speziellen Problemen angemessene Formeln. Bei Luftgeschwindigkeiten v unter 5 m/sek parallel zur Wärmeübergangsfläche gilt nach Jürges [4]

$$(3) \quad \alpha = 5,3 + 3,6 \cdot v \text{ (kcal/m}^2 \cdot h \cdot ^\circ\text{C)}$$

Der von Angström bei $v = 2,8$ m/sek gefundene Wert $\alpha = 0,03 \text{ cal/cm}^2 \cdot \text{min} \cdot ^\circ\text{C}$ stimmt ungefähr mit dem nach Formel (3) berechneten überein ($\alpha = 0,026 \text{ cal/cm}^2 \cdot \text{min} \cdot ^\circ\text{C}$). Für grössenordnungsmässige Abschätzungen des Wärmeüberganges haben wir daher diese Gleichung gelegentlich beigezogen. Weitere Untersuchungen in dieser Richtung, vor allem direkte experimentelle Bestimmungen des Wärmeüberganges, wären sehr wünschenswert. Zum Wärmeübergang zwischen Luft und Schneedecke ist noch ein gewisser Wärmetransport hinzuzurechnen, der mit dem *Luftaustausch* verbunden ist. Wir werden diese Frage noch im Zusammenhang mit der Wärmeübertragung im Inneren der Schneedecke kurz streifen.

c) *Wärmeübergang vom Boden in die Schneedecke.*

Der Wärmestrom vom Boden in die Schneedecke kann aus dem Temperaturgradienten in den bodennahen Schichten ermittelt werden, falls die Wärmeleit-

fähigkeit des betreffenden Schnees bekannt ist. Dabei muss allerdings vorausgesetzt werden, dass die Bodenwärme ausschliesslich durch Wärmeleitung und nicht durch Konvektion der Porenluft abgeführt wird — es sei denn der verwendete Wärmeleitkoeffizient schliesse die Konvektion bereits ein. Unter dieser Annahme berechneten wir aus den Temperaturprofilen des Versuchsfeldes Weissfluhjoch im Winter 1945/46 den in folgendem Diagramm (Bild 3) wiedergegebenen Wärmestrom. Die zu

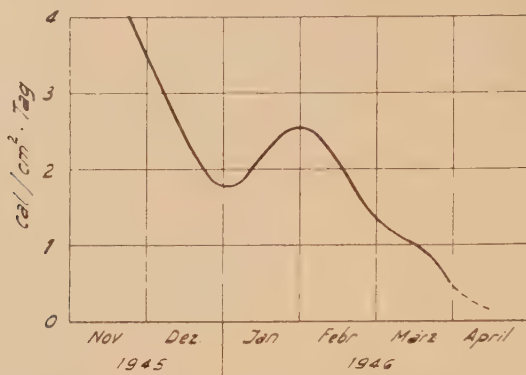


BILD 3 Wärmestrom vom Boden in die Schneedecke, berechnet aus Temperaturgradienten.
($\lambda_s = 5 \cdot 10^{-4} \text{ cal} \cdot \text{cm}^{-1} \cdot \text{sek}^{-1} \cdot \text{grad}^{-1}$) Versuchsfeld Weissfluhjoch.

Beginn des Winters starke Wärmeabgabe erklärt sich aus der noch hohen Bodentemperatur und der geringen Schneebedeckung. Das Zwischenmaximum anfangs Februar ist den tiefen Januartemperaturen zuzuschreiben. Anfangs April hört die Wärmeabgabe des Bodens natürlich nicht auf; die Wärme dringt jedoch nicht mehr in die aufgetaute Schneedecke ein, sondern wird an der Schneeoberfläche in Schmelzwasser umgesetzt. Dieser Beitrag ist aber sehr klein, von der Grössenordnung 0,01 mm H₂O pro Tag. Erst wenn die Einstrahlung durch den Schnee hindurch den Boden aufzuheizen vermag, wird der Abbau der Schneedecke vom Boden her merklich.

d) Verdunstung und Kondensation.

Die Wärmetönung beim Wasserdampfaustausch zwischen Schnee und Atmosphäre beträgt für 1 mm Wasseraequivalent 67,6 cal/cm². Diese Wärmemenge wird teils der Luft, zur Hauptsache aber der obersten Schneeschicht entzogen und trägt, wie die Ausstrahlung, zur Unterkühlung der Schneeoberfläche bei. Der Wasserdampfaustausch interessiert uns aber nicht nur wegen seiner Bedeutung für den Wärmehaushalt, sondern ebenso sehr wegen des mengenmässigen Umsatzes. Seit Jahrzehnten ist die Bestimmung der Verdunstungsgrösse eine wichtige Aufgabe der Hydrologie.

Der Wasserdampfaustausch hängt bei klarem Wetter in hohem Mass von der Hangexposition und vor allem von der Sonnenexposition ab. In Sonnenlage ist die Verdunstung stets viel grösser als am Schatten solange die Schneeoberfläche zeitweise oder dauernd negative Temperaturen aufweist. Wir fanden z.B. vom 14. April 1947 0915 bis 15. April 0915 eine Verdunstung von

1,3 mm H₂O Tag in Sonnenlage
0,6 mm H₂O Tag in Schattenlage

(Luftfeuchtigkeit : 55 %, Windgeschw. : 1,5 m/sek., Globahlstrahlung : 740 cal/cm², Minimale Schneeoberflächentemperatur : — 4° C)

Für die in Sonnenlage verdunsteten 1,3 mm wurde eine Energie von 88 cal/cm² d.h. fast 1/3 der Einstrahlung (Albedo = 0,6), verzehrt. Gelegentlich konnten wir beobachten, dass in Sonnenlage Schnee verdunstete, während sich in einem abgeschatteten Feld unmittelbar daneben gleichzeitig Wasserdampf auf der Schneeoberfläche kondensierte. Nebenbei sei bemerkt, dass wir sowohl im Feld wie im Laboratorium stets grössere Verdunstungsbeträge über Eisflächen gemessen haben als über Schneeflächen, zweifellos wegen der grösseren Unterkühlungsfähigkeit der Schneeoberfläche.

Als Gesamtbilanz des Wasserdampfaustausches im horizontalen Versuchsfeld Weissfluhjoch haben wir gesamthaft für die Monate Januar-April 1947 auf Grund unserer Messreihen folgende Werte abgeschätzt :

Verdunstung in Sonnenlage	36 mm H ₂ O
» » Schattenlage	10 mm H ₂ O

Der Sonnenwert, der zu ca. 3/4 auf den April entfällt, entspricht ungefähr einem Anteil von 6 % des Höchstwasserwertes der Schneedecke 1946/47. Bevor also die eigentliche Schneeschmelze einsetzt, haben Verdunstung und Kondensation den Schneeabbau eingeleitet und dabei eine Differenzierung nach Sonnen- und Schattenlage herbeigeführt, die dann in der Schmelzperiode noch viel stärker in Erscheinung tritt. Die erwähnten Zahlen beziehen sich auf das horizontale Versuchsfeld. Für sonnenwärts geneigte Schneeflächen wird die spezifische Einstrahlung noch grösser und entsprechend steigen auch die Verdunstungsbeträge.

3. Wärmeübertragung im Innern der Schneedecke

a) Die Wärmeleitung in Trockenschnee.

Wärmeleitkoeffizienten von Trockenschnee sind von verschiedenen Forschern gemessen und in Formeln gefasst worden, welche den Zusammenhang mit dem Raumgewicht wiedergeben. Im Diagramm Bild 4 sind solche Funktionen dargestellt. Die Uebereinstimmung der verschiedenen Kurven ist bis zu den mittleren Raum-

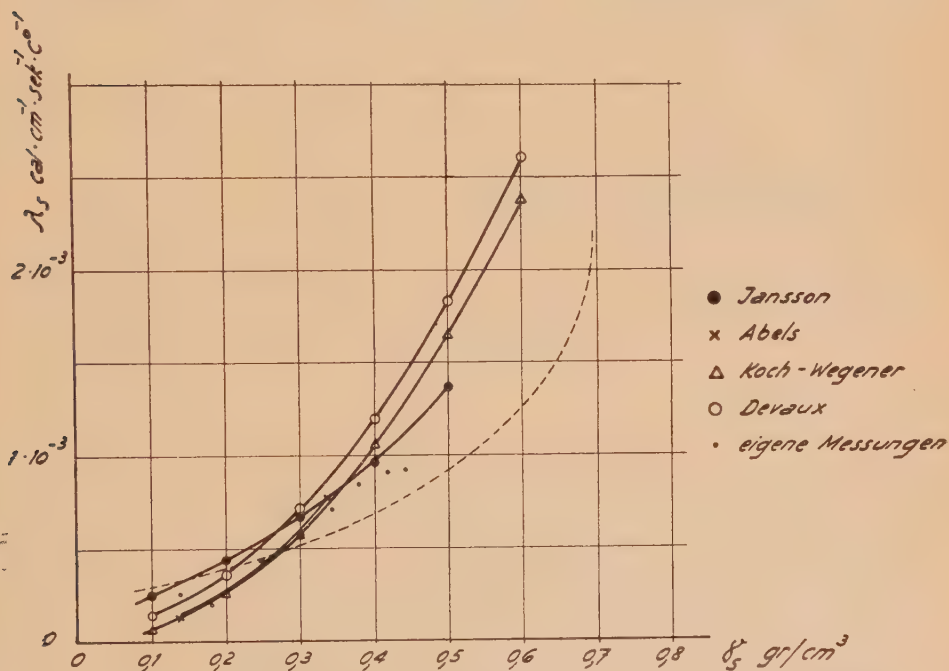


BILD 4 Wärmeleitkoeffizienten und Raumgewicht von Schnee nach Jansson (5) Abels (6) Koch-Wegener (8) und Devaux (7). Gestrichelt: Für Eiskugel-Struktur berechnete Kurve.

gewichten nicht schlecht. In zweifacher Hinsicht wären noch Ergänzungen erwünscht, nämlich Angaben über den Einfluss der Korngrösse auf die Wärmeleitung, und über die Bedeutung der Luftkonvektion. Wir sind mit solchen Untersuchungen beschäftigt und können die gewünschten Ergänzungen noch nicht zahlenmässig geben. Die allgemeine Zunahme der Wärmeleitfähigkeit mit dem Raumgewicht deutet aber darauf hin, dass die Wärmeübertragung jedenfalls zum grösseren Teil den Weg über das Korngerüst nimmt. Berechnet man Wärmeleitkoeffizienten für

idealisierte Eiskugelstrukturen vom Raumgewicht des Schnees, kommt man bereits in die Grössenordnung der wirklich an Schnee gemessenen Wärmeleitkoeffizienten. Die dazu notwendigen Annahmen über Zahl und Grösse der Kornberührungsflächen können den natürlichen Verhältnissen im Kornverband auf Grund direkter Beobachtungen entnommen werden. (Schneedünnschliffe). Schliesslich haben unsere bisherigen Wärmeleitversuche an kugelförmigen im Zentrum geheizten Schneekörpern nur eine schwache Richtungsabhängigkeit der Wärmeausbreitung erkennen lassen. Unter Berücksichtigung der Porosität und Luftdurchlässigkeit von Schnee kann der durch Diffusion und thermische Konvektion verursachte Wärmetransport in der Porenluft etwa zu 10-50 % der gesamten Wärmeleitung veranschlagt werden. Mit zunehmendem Raumgewicht geht er natürlich zurück.

b) *Eindringen von periodischen Temperaturschwankungen und von Temperatursprüngen in Trockenschnee* (Ohne Berücksichtigung der Strahlung)

Die eigentliche Schneeschmelze wird im Alpengebiet gewöhnlich durch eine längere Periode starker *täglicher Schwankungen der Lufttemperatur* eingeleitet. Die Tiefenwirkung und zeitliche Verzögerung beim Eindringen dieser Schwankungen bestimmen die erste Phase des Schneedeckenabbaues. Wir haben versucht sowohl von der rechnerischen als auch von der experimentellen Seite her Einblick in die Temperaturbewegung zu erhalten, ähnlich, wie dies Hughes und Seligman [9] für die Jahresschwankungen der Firntemperaturen gemacht haben. Die rechnerische Behandlung kann der Spezialliteratur über Wärmeleitung entnommen werden (z.B. Groeber-Erk : Die Grundgesetze der Wärmeübertragung). Eine sinusförmige oberflächliche Temperaturschwankung vom Betrag $\pm \delta_0$ um die mittlere Temperatur δ_m führt zu folgenden Temperaturverteilung in einer unbegrenzt tief gedachten Schneedecke :

$$(4) \quad \delta(z,t) = \delta_m + \delta_0 \cdot e^{-z\sqrt{\frac{\pi}{aT}}} \cdot \cos\left(z\sqrt{\frac{\pi}{aT}} - \frac{z\pi}{T} t\right)$$

z = Tiefe unter der Schneeoberfläche im cm.

a = Temperaturleitfähigkeit von Schnee

T = Periodenlänge der oberflächlichen Temperaturschwankung.

t = Zeit

Die Temperaturamplitude klingt also mit der Tiefe exponentiell ab und erleidet eine Phasenverzögerung.

$$t_z = \frac{z}{2} \sqrt{\frac{T}{\pi a}}$$

Ein praktisches Beispiel :

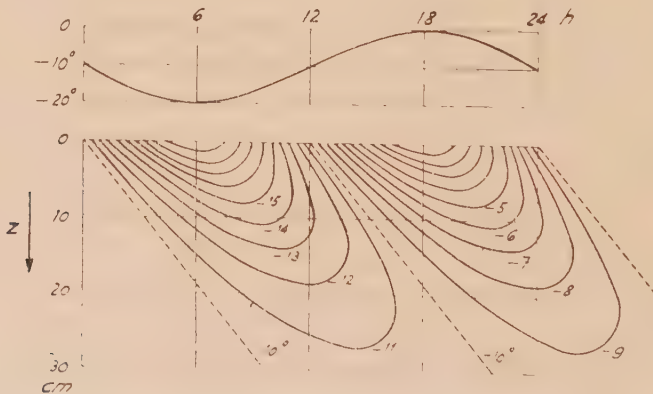


BILD 5 Das Eindringen von periodischen Schwankungen der Obefflächentemperatur in die Schneedecke (Tagesschwankungen)
Oben : Sinusförmiger Verlauf der Obefflächentemperatur $\delta(t)$
Unten : Zeitlicher Isothermenverlauf (S = Schneeoberfläche)

Eine oberflächliche tägliche Temperaturschwankung von $\pm 10^\circ$ erzeugt Temperaturamplituden von

$\pm 4,3^\circ$	in	10 cm	Tiefe
$\pm 0,14^\circ$	»	80 »	»
$\pm 0,02$	»	100 »	»

$$(a = \frac{\lambda_s}{c_s \gamma_s} = 5.10^{-3} \text{ cm}^2/\text{sek gesetzt})$$

Die Fortpflanzungsgeschwindigkeit der Temperaturwelle beträgt dabei 73 cm/Tag. Bild 5 zeigt den Isothermenverlauf für das berechnete Beispiel in einer Oberflächenschicht von 30 cm. Die Tiefenwirkung Tagesschwankungen ist also klein und nur bis in eine Tiefe von ca. 50 cm direkt messbar. Laufende Temperaturregistrierungen in verschiedenen Schneeschichten ergaben grundsätzlich dasselbe Bild.

Schneller und nachhaltiger dringt ein *aperiodischer Temperatursprung* in die Schneedecke. Die Rechnung ergibt für einen Temperatursprung von 20°C . den in Bild 6 dargestellten Isothermenverlauf.

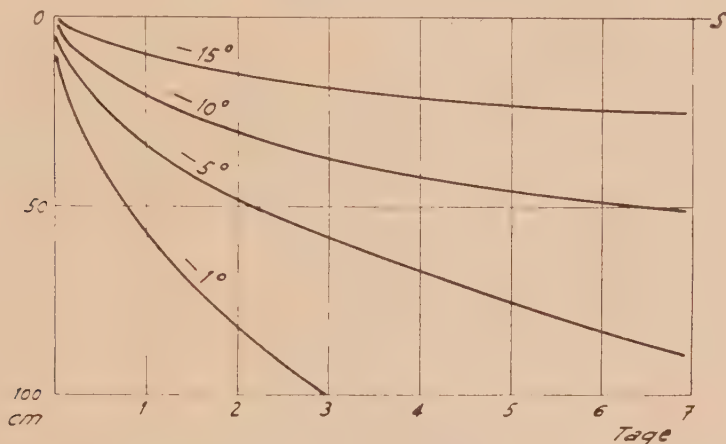


BILD 6 Eindringen eines Temperatursprungs in die Schneedecke.

Anfangsbedingung : Schnee- und Lufttemperatur = 0° (trocken). Nach $t = 0$ Schneeoberflächentemperatur = -20° . (s = Schneeoberfläche, z = Tiefe unter der Oberfläche).

c) Wärmeübertragung in Nass-Schnee.

Die Wärmeübertragung kann in Nass-Schnee, einem isothermen System, nicht auf dem Weg der Wärmeleitung erfolgen. Es ist daher nicht gut verständlich, wenn in der Literatur Wärmeleitkoeffizienten für Nass-Schnee angegeben werden. R.E. Horton [10] hat gezeigt, wie durch Schmelzwasser ein gewisser Wärmetransport von der Schneeoberfläche nach der Tiefe bewerkstelligt wird. Solange die Schneedecke im Innern noch Minustemperaturen aufweist, wird das Schmelzwasser an der momentanen Nullgrad-Grenze aufgefangen und gefroren. Diese wird dabei durch die freiwerdende Erstarrungswärme weiter nach unten verschoben. Sobald die Schneedecke ganz aufgetaut ist, wird die Rolle des Schmelzwassers als Wärmeträger für tiefere Schichten sicher unbedeutend. Schmelzwasser und auch flüssiger Niederschlag verlieren ihre Schmelzkraft bereits in Berührung mit den obersten Schneelagen. Etwas tiefer reicht sie, wenn sich das Wasser in einzelnen Kanälen sammeln kann (Regenrinnen). Jedenfalls stellen wir immer wieder fest, dass der Wasserwert von Schichten, die sich mehr als ca. 50 cm unter der Oberfläche befinden, während der Abbauzeit nur wenig abnimmt.

Das Wiedergefrieren einer aufgetauten Schneedecke geht in ähnlicher Weise vor sich wie das Eindringen von Frost in feuchten Boden und lässt sich rechnerisch behandeln. Bei konstanter negativer Schneeoberflächentemperatur hängt die Tiefe z der Nullgradgrenze wie folgt von der Zeit ab :

$$z = A \cdot \sqrt{t}$$

Die Grösse A wird bestimmt durch die Oberflächentemperatur, den Wassergehalt des Nass-Schnees, das Raumgewicht und durch die Temperaturleitfähigkeit des wiedergefrorenen Schnees. Für Schnee von einer Feuchtigkeit von $F = 10\%$ und einer Temperaturleitfähigkeit von $a = 5 \cdot 10^{-3} \text{ cm}^2/\text{sek}$ berechnet man bei einer Oberflächentemperatur von -5° : $A = 0,053 \text{ cm}/\text{sek}^{-1/2}$. Die Frostgrenze dringt mit diesem Wert von A in den ersten 24 Std. nach dem Gefrieren der Oberfläche in eine Tiefe von ca. 15 cm vor. Laboratoriumsversuche ergaben ein etwas schnelleres Vordringen der Frostgrenze ($z = A \cdot t^{2/3}$), wahrscheinlich wegen der Luftzirkulation.

4. Wasserspeicherung in Schnee und Abflussverzögerung

a) Definition der Schneefeuchtigkeit.

Unter der *absoluten Feuchtigkeit* F soll der Gewichtsanteil flüssigen Wassers von Nass-Schnee verstanden werden, ausgedrückt in % also

$$F = 100 \frac{W}{E + W} \%$$

W = Gewichtsanteil Wasser

E = Gewichtsanteil Eis.

Ferner verwenden wir gelegentlich folgende Begriffe: Die *Sättigungsfuchtigkeit* F_s ; das ist die Feuchtigkeit bei vollständiger Wasserfüllung des Porenraumes, also die höchstmögliche absolute Feuchtigkeit. Als *relative Feuchtigkeit* f bezeichnen wir das Verhältnis $100 \cdot \frac{F}{F_s}$ d.h. den prozentualen Anteil der Porenfüllung.

Die *Gleichgewichtsfeuchtigkeit* wird später erläutert.

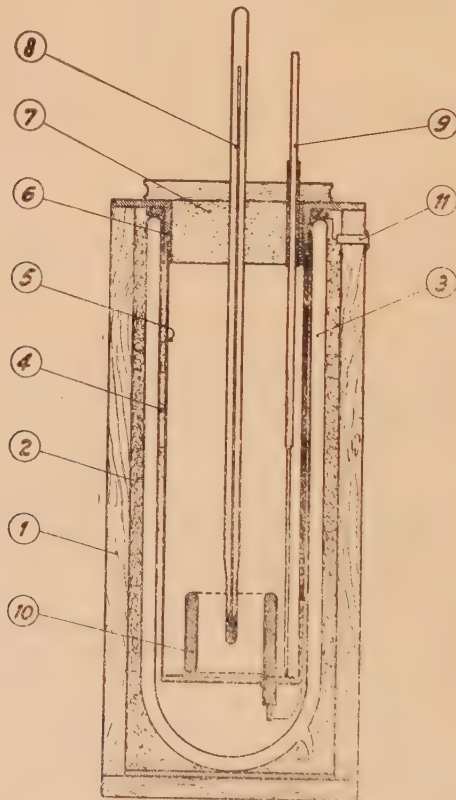


BILD 7 Kalorimeter zur Bestimmung des Wassergehaltes von Nass-Schnee.

1 Holzkasten, 2 Glaswolle, 3 Dewar-Gefäss, 4 Luftraum, 5 Kupfergefäss, 6 Presstoffring, 7 Deckel
8 Thermometer, 9 Rührer, 10 Heizelement, 11 Elektr. Anschluss.

b) Messung der Schneefeuchtigkeit.

Als beste Methode zur Bestimmung der abs. Schneefeuchtigkeit hat sich bisher die kalorimetrische Messung des Schmelzwärmebedarfes von Nass-Schnee erwiesen. Eine gewogene Probe Nass-Schnee wird in einem Kalorimeter (Bild 7) durch einen elektrischen Heizkörper geschmolzen, indem eine bestimmte, gerade ausreichende Wärmemenge zugeführt wird. Aus der Temperatur des Schmelzwassers, dem Gewicht der Schneeprobe, der aufgewendeten Energie und der Wärmekapazität des Kalorimeters lässt sich die abs. Feuchtigkeit mit einem maximalen Fehler von $\Delta F = \pm 1,5 \%$ berechnen. Zur Messung der Heizenergie bedient man sich eines Wattmeters, oder man heizt das Schmelzwasser nachträglich noch während 1 - 2 Minuten auf und ermittelt den Heizwert des Stromes kalorimetrisch.

Weniger genau, dafür unabhängig von einem elektrischen Anschluss wird die Feuchtemessung, wenn man die Schmelzwärme in der Form von heissem Wasser zuführt.

Die Messungen im Laboratorium ergaben, dass sich im Nass-Schnee eine bestimmte absolute Feuchtigkeit einstellt, die nur von der Schneestruktur (Korngrösse und Raumgewicht) abhängt. Wird Nass-Schnee mit Eiswasser übergossen, so fliesst das überschüssige Wasser ab, und man misst nach einiger Zeit wieder die ursprüngliche Feuchtigkeit, die wir als *Gleichgewichtsfeuchtigkeit* bezeichnen.

Für die verschiedenen Schneearten fanden wir folgende Abstufung der Gleichgewichtsfeuchtigkeit :

Schneeart	Raumgewicht	F
Neuschnee	300 - 400 kg/m ³	20 - 30 %
Altschnee feinkörnig	400 - 450 »	10 - 20 %
Altschnee grobkörnig	450 - 500 »	5 - 10 %

Wird an der Basis der Schneeproben der Wasserabfluss verhindert, bildet sich über dem Wasserspiegel eine mit Wasser völlig gesättigte Zone von 3-8 cm Dicke, je nach Kornstruktur (siehe Bild 8). In der natürlichen Schneedecke fanden wir am 1. Mai 1946 das in Bild 9 wiedergegebene Feuchtigkeitsprofil.

c) Wasserdurchlässigkeit von Schnee und Verzögerung des Wasserabflusses.

Die Wasserdurchlässigkeit kann auf verschiedene Arten definiert werden :

1. als Wassermenge, welche in der Zeiteinheit unter einem bestimmten Druckgefälle durch einen Schneequerschnitt fliesst.
2. als Versickerungsgeschwindigkeit von Regen- oder Schmelzwasser in Schnee.

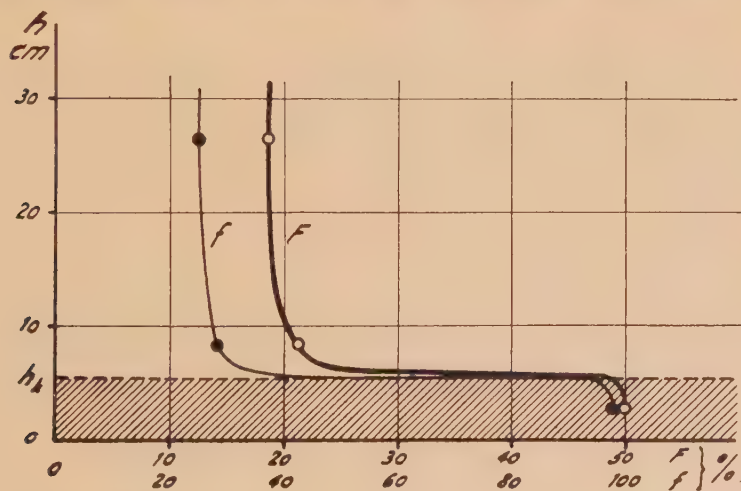


BILD 8 Feuchtigkeit von Nass-Schnee in Berührung mit einem Wasserspiegel (Feinkörniger Altschnee).
 Absolute (F) und relative (f) Feuchtigkeit.
 h = Höhe über dem Wasserspiegel.
 h_k = Höhe der kapillar gesättigten Zone.

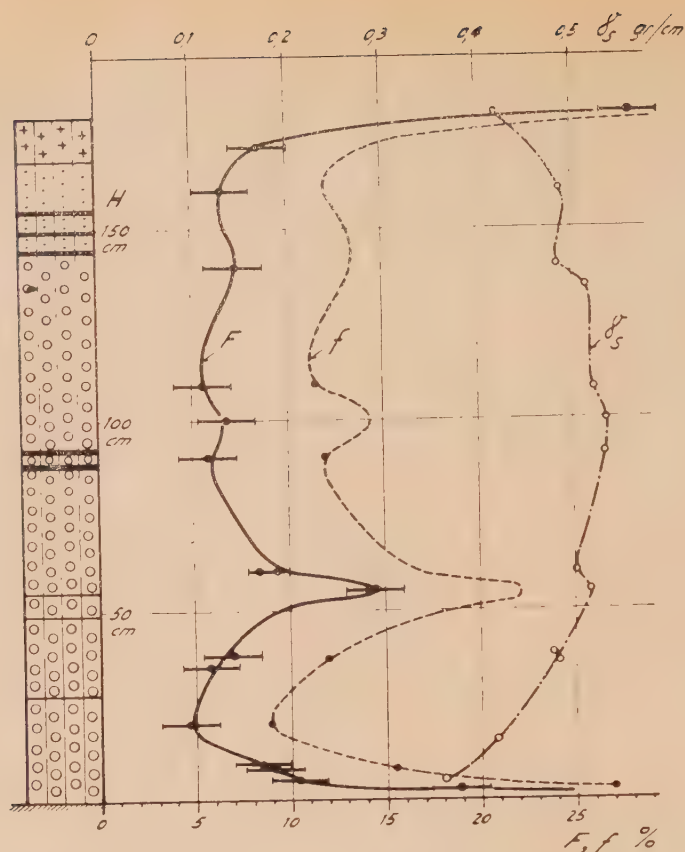


BILD 9. Feuchtigkeitsverhältnisse und Raumgewicht in der natürlichen Schneedecke. (1.5.46. Versuchsfeld Weissfluhjoch).

H = Höhe über dem Boden. γ_s = Raumgewicht.

Im ersten Fall erfasst man die Verhältnisse bei Sättigungsfeuchtigkeit, z.B. an der Basis einer horizontalen Schneedecke; im zweiten Fall bei einem Wassergehalt zwischen Gleichgewichtsfeuchtigkeit und Sättigungswert, also zwischen der Snowoberfläche und der gesättigten Basiszone. Beide Formen der Wasserführung sind an der Entwässerung der Schneedecke beteiligt.

Zur Bestimmung der Wasserdurchlässigkeit bei Sättigungsfeuchtigkeit haben wir direkte Messungen vorgenommen und zum Vergleich auch Luftdurchlässigkeitswerte beigezogen. Die letzteren können mit dem Quotienten der Zähigkeitswerte von Wasser und Luft ($\frac{\eta_w}{\eta_l} = 104$ bei 0°) auf Wasserdurchlässigkeitswerte umgerechnet werden. Bei diesem Vergleich ist allerdings zu berücksichtigen, dass sich mit der Durchnässung die ganze Schneestruktur meistens etwas kontrahiert. Nachfolgend geben wir einige Durchflusszahlen D in cm^3/sek durch einen Schneequerschnitt von 1 cm^2 bei einem Druckgefäll vom 1 dyn/cm^3 . (D_I direkt gemessen, D_{II} aus Luftdurchlässigkeit umgerechnet).

Korngrösse (Altschnee)	Raumgewicht (feucht)	D_I	D_{II}
feinkörnig ($d < 1 \text{ mm}$)	0,352 gr/cm^3	0,58 10^{-3}	0,53. 10^{-3}
mittleres Korn ($0,8 < d < 1,5 \text{ mm}$)	0,382 „	1,22	1,27
grobkörnig ($d > 2 \text{ mm}$)	0,385 „	2,28	2,20
„ ($d > 2 \text{ mm}$)	0,340 „	4,30	4,05

Die *Versickerungsgeschwindigkeit* in vertikaler Richtung wurde an zylindrischen Schneeproben von 12 cm Durchmesser und ca. 35 cm Länge gemessen. In die obere Stirnseite wurde Eiswasser in konstantem Zustrom eingegossen, und bei seinem Austritt auf der Unterseite floss es direkt in eine Waagschale. Kleinere Zuflussmengen haben wir erzeugt, indem wir die obere Seite des Schneezylinders in Berührung brachten mit einer schwach geheizten elektrischen Platte.

Die Proben müssen bei diesen Versuchen bereits durchgehende Gleichgewichtsfeuchtigkeit besitzen und zudem an der Basis gesättigt sein. Drei Zahlenbeispiele :

S	Z (mm /Min)	V (cm /Min)	t_v (Min)
feinkörnig (1 mm)	0,38	1,47	68
$\gamma_s = 451 \text{ kg/m}^3$	0,92	2,78	36
grobkörnig (2 mm)	0,47	1,25	80
$\gamma_s = 503 \text{ kg/m}^3$	0,92	2,56	39
grobkörnig (2,5 mm)	0,36	1,61	62
$\gamma_s = 470 \text{ kg/m}^2$	1,2	3,45	29

(S = Schneeart, Z = Wasserzufluss, V = Versickerungsgeschwindigkeit
 t_v = Abflussverzögerung für 1 m Schnee.)

Wenn der Zufluss unterbrochen wird, lässt der Abfluss bereits vor der Zeit t_v nach und klingt dann während einer wesentlich längeren Zeit allmählich aus. (Bild 10)

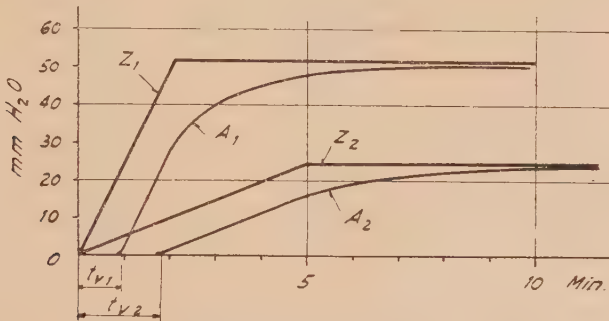


BILD 10 Versickerung von Eiswasser in Schnee in vertikaler Richtung. Versickerungsstrecke = 100 cm.
Z = Zeitlicher Verlauf des Zuflusses an der Oberfläche.
A = Zeitlicher Verlauf des Abflusses an der Basis.
 t_v = Verzögerung der Abflusses.
 $\frac{100}{t_v}$ = Versickerungsgeschwindigkeit

Einige weitere Versuche wurden ferner unternommen, um Aufschluss zu erhalten über den Wasserabfluss längs der Basis einer geneigten Schneedecke. Zu diesem Zweck wurden Schneequader von ca. 1 m Länge, die sich auf einer um 22° geneigten Unterlage befanden, am höher liegenden Ende in konstantem Zustrom mit Eiswasser begossen. Am tieferliegenden Ende floss das Wasser in ähnlicher Weise wie bei den Versuchen mit vertikaler Versickerung aus. Wie erwartet, wurde aber längs der unteren Grenzschicht eine grössere Versickerungsgeschwindigkeit gemessen. Sie betrug beispielsweise 22 cm/Min bei einem Zufluss von 0,6 Liter/Min auf einer Probenbreite von 100 cm. (Raumgewichte des Schnees : 0,382 gr/cm³, Korngrösse : 0,5 - 1,2 mm) 45 Min nachdem der Zufluss abgestellt war, also ca. nach der 10 fachen Verzögerungszeit, gab der Schnee noch ca. 1/50 der maximalen Abflussmenge ab.

Mit Hilfe solcher experimenteller Unterlagen kann man für idealisierte Verhältnisse den zeitlichen Abfluss aus einer geneigten Schneedecke bei bekannter Schmelzwasser- oder Niederschlagsmenge berechnen, ähnlich wie dies Horton angegeben hat.

5. Die alpine Schneedecke während der Schneeschmelze

Die einzeln besprochenen Vorgänge der Wärmeübertragung und der Wasserbewegung im Schnee spielen sich in der Natur oft gleichzeitig ab und beeinflussen sich gegenseitig. An Hand eines Beispiels des alpinen Schneedeckenabbaues sollen nun die Einzelerscheinungen in ihrem Zusammenwirken betrachtet werden. Abschliessend wollen wir versuchen diejenigen Bedingungen festzuhalten, welche zu abnormalen und gefährlichen Wasserabflussmengen führen können.

Die Schneedecke möge anfangs April ihren Höchststand von ca. 3 m aufweisen, gemessen auf einem horizontalen, sonnenexponierten Feld. Im stark gegliederten Gelände hat aber die Windwirkung schon bei der Ablagerung grosse örtliche Unterschiede in der Schneehöhe verursacht. Auf engem Raum wird man Schneeanhäufungen von über 5 m und Stellen von nur ganz geringer Schneeüberdeckung finden. Die Verdunstung hat überdies bereits eine gewisse Differenzierung in den Schneemengen zwischen Schatten- und Sonnenhängen hervorgerufen. Unter den täglichen Schwankungen der Lufttemperatur und der Strahlung, dringen Wärme- und Kälte- wellen ins Schneeeinnere ein, wobei sich die Mitteltemperaturen immer mehr dem Nullpunkt nähern. (Bild 11a) Lange bevor der Kältegehalt der Schneedecke aufgezehrt ist, beobachtet man das erste vorübergehende Auftauen der Schneeoberfläche. Die zunächst geringen Schmelzwassermenge vermag noch nicht in die Tiefe zu dringen. Gleich bei Sonnenuntergang gefriert sie jeweils wieder und verursacht einen Sonnenharsch. Von Tag zu Tag wird aber die Tiefe der Durchnässung stärker.

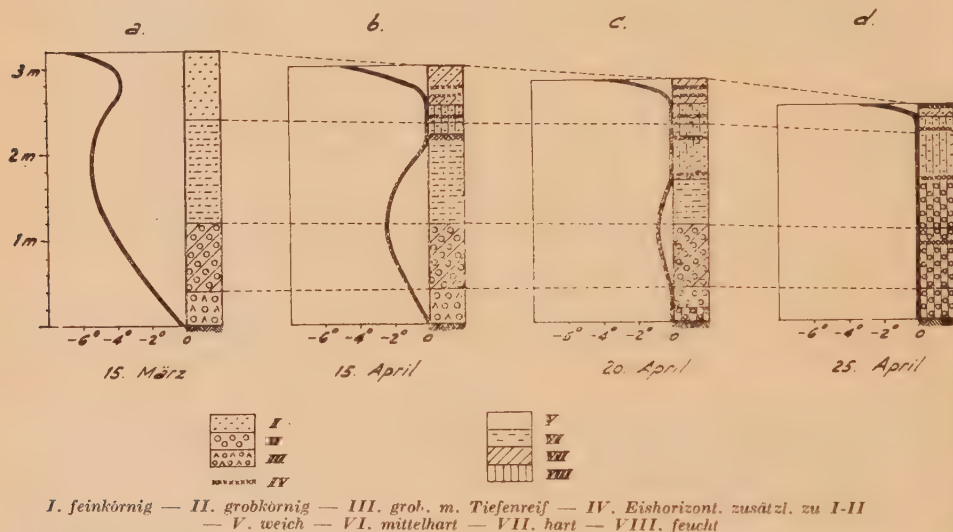


BILD 11 Temperatursverhältnisse in der Schneedecke von der beginnenden Aufwärmung bis zur völligen Durchnässung. (schematisch) Zustand am Morgen der betr. Tage.

Es bilden sich an der Nullgradgrenze nicht weit unter der Schneeoberfläche die ersten Eislamellen. In der Nacht steigt die Nullgradgrenze zurück und vereinigt sich anfänglich wieder mit der von der Oberfläche eindringenden Frostgrenze. Eines Tages bleibt eine Nass-Schneezone bestehen (Bild 11b). Nun wandert die Grenze zwischen feuchtem und trockenem Schnee immer schneller in die Tiefe, und die Eislamellen folgen sich in immer grösseren Abständen (11c). Während der Aufwärmung von oben vermag die Bodenwärme den Grund des Profils aufzutauen. Gegen Ende April finden wir dann tagsüber die ganze Schneedecke aufgeweicht, und wenige Tage nachdem Auftauen des ganzen Profils wird auch die Sättigungsfuchtigkeit durchgehend vorhanden sein (11d). Noch immer verzögern aber die natürlichen Kälterückfälle den eigentlichen oberflächlichen Abbau der Schneedecke. Bis zu diesem Stadium zeigen die geschilderten Vorgänge eine enorme zeitliche Verschiebung nach Höhenlage und Hangexposition. So wird z.B. die durchgehende Durchnässung der Schneedecke auf Büschalp, einer 700 m tiefer gelegenen Bergterrasse, ca. 3 Wochen früher erreicht als auf Weissfluhjoch.

Dieser für Schönwetterperioden zutreffende Auftauvorgang wird bei anderen Witterungsverhältnissen in verzögerndem oder beschleunigendem Sinn abgewandelt. Es schieben sich Kälteperioden ein und bringen neuen Schneefall, oder dann überfluten Warmluftmassen das Land und zehren den Kältevorrat der Schneedecke in wenigen Tagen auf, und zwar ohne grosse Unterschiede zwischen Sonnen- und Schattenlage. Im Alpengebiet ist es der bekannte Föhn, der dem Schnee am meisten zusetzt, vor allem in den Tälern.

Dort, wo die Schneedecke sich im Zustand der Sättigungsfeuchtigkeit befindet, kann der Abfluss einsetzen. Ein Teil des Wasser folgt den Eisschichten, wird aber im allgemeinen bald zum Boden absickern. Die seitliche Verzögerung von einer bis mehreren Stunden bei der vertikalen Versickerung fällt bei der Entwässerung eines grösseren Gebietes weniger ins Gewicht als die beschränkte Strömungsgeschwindigkeit längs der Bodenoberfläche und im Boden selbst.

Schicht um Schicht wird nun die Schneedecke von oben her abgetragen. Die innere Setzung, die während des Hochwinters mit stetig abnehmender Geschwindigkeit vor sich ging, beschleunigt sich in der Abbauperiode etwas. Vor allem verdichten sich die untersten aus lockerem, grobkörnigem Schwimmschnee bestehenden Schichten. Bild 1 zeigt die Lage einzelner innerer Schichten, gemessen mit elektrischen Pegeln. Erst kurz bevor die einzelnen Pegel an die Oberfläche gelangen, kann man ein starkes Schmelzen ihrer Umgebung erkennen. Die Endphase des Abbaues geht sehr schnell vor sich. Von oben, von unten und im Innern selbst werden die letzten 50 cm Schnee angegriffen. Die Schmelzwasserproduktion ist wegen den höheren Mitteltemperaturen und der stets zunehmenden Einstrahlung intensiv. Das Material wird, wie man sagt, faul, und ahnungslose Skifahrer brechen gelegentlich mit-samt ihren Brettern ein. Infolge der inneren Auflösung wird die Wasserspeicherungsfähigkeit des Schnees etwas abnehmen, sodass mit dem frischen Schmelzwasser noch Speicherwasser zusätzlich abfließt. Wir halten diesen Zuschuss jedoch in den geschilderten Verhältnissen für unbedeutend. Wenn die letzten Schneereste auf unserem Versuchsfeld bereits verschwunden sind, finden sich in der Umgebung da und dort noch metertiefe Anhäufungen und legen Zeugnis ab von der Ungleichheit der Ablagerung und des Abbaus. Ihre Liquidierung kann sich noch bis in den Spätsommer hinziehen.

Fragt man sich nun, unter welchen Bedingungen eine Schneedecke so schnell aufgelöst werden kann, dass Hochwassergefahr zu befürchten ist, dann gibt das erläuterte Beispiel indirekt gewisse Hinweise. Es lässt einerseits *topographische* und anderseits *meteorologische Faktoren* erkennen, welche der Hochwassergefahr Vorschub leisten.

Topographische Faktoren sind : Offenes, wenig gegliedertes Gelände mit einheitlicher Besonnung und nur geringen Höhenunterschieden. Die Schneedecke ist überall und gleichzeitig dem Abschmelzen ausgesetzt. Ferner sind hier zu erwähnen : Schlechte Wasserdurchlässigkeit und -Speicherungsfähigkeit des Bodens,

Meteorologische Faktoren sind : Anhaltender Zustrom warmer, feuchter Luft, besonders wenn verbunden mit grosser Windgeschwindigkeit und mit Regenfällen. Auch ohne starke Einstrahlung wird dabei viel Schmelzwasser erzeugt, da die kühlende Verdunstung ausbleibt und die Schattenhänge in gleichem Mass wie die Sonnenhänge abgebaut werden. Trockene Schönwetterperioden sind im Gebirge erst in zweiter Linie zu nennen.

Besonders gefährlich werden sich Kälterückfälle mit grösseren Schneefällen in einer normalerweise warmen Jahreszeit auswirken. In der zu warmen Umgebung wird der Neuschnee schnell aufgezehrt und setzt seinen Wassergehalt plötzlich frei.

Wenn man von lokalen Ereignissen, wie Gewässerstauungen durch Lawinen mit nachfolgendem Wasserdurchbruch und dergleichen absieht, wird man den Schnee selbst nicht der aktiven Mitverantwortung für Hochwassergefahren bezieh-tigen können. In unseren alpinen regionen ist jedenfalls dafür gesorgt, dass die Schnee-decke ihren Wassergehalt in dosierten, auf Monate verteilten Mengen abgibt.

LITERATUR

- (1) P. NIGGLI, H.BADER, R. HAEFELI, E. BUCHER, J. NEHER, O. ECKEL und Ch. THAMS. Der Schnee und seine Metamorphose (V Teil). Beitrage zur Geologie der Schweiz. Geotechnische Serie, Hydrologie Lieferung 3.
- (2) DEFANT A. Met. Z. 36 (1919) 153.
- (3) NIEDEEDORFER E. : Met. Z. 50 (1933) 201.
- (4) JÜRGES W. : Der Wärmeübergang an einer ebenen Wand. Beiheft z. Gesundheitsing. Reihe 1 No. 19, 1923.
- (5) JANSSON M. : Ueber die Wärmeleitfähigkeit des Schnees :
Öfversigt Kongl. Vetenskaps. Akad. Forhandlingar Stockholm 58 (1901).
- (6) ABELS H. : Rep. f. Meteor. Petersburg 16 (1892) No.1.
- (7) DEVAUX J. : L'économie radiothermique des champs de
neige et des glaciers. Am. Phys. 20 (1933) No. 2.
- (8) KOCH WEGENER : Meddeleser om Groenland 75. 191 (1930), Zitiert nach (9).
- (9) HUGHES T. P. und SELIGMAN G. : Monthly Notices of R. A. S, Geophysical
Supplement 1939 Dec.
- (10) HORTON R. E. : Transactions of the Meeting of the Interna. Comm. of snow
and of Glaciers (Edinburgh 1936).
- (11) H. N. SVERDRUP : Geografiska Ann. XVII (1935) Stockholm.

} Zitiert nach (1)

8) THE COOPERATIVE SNOW INVESTIGATIONS PROGRAM. ITS OBJECTIVES AND OPERATIONS

by

F. L. RHODES and W. T. WILSON (U.S.A.)

INTRODUCTION

In recognition of the influence of snow on the planning, design, and operation of flood control and multiple purpose projects in the mountain regions of the United States there was initiated in 1945 by the Weather Bureau and the Civil Works Division of the Corps of Engineers, on a cooperative basis, a comprehensive snow investigations program to :

1. Determine a practicable and reliable method of evaluating the maximum streamflow which may be produced by snowmelt and rain ;
2. Develop a practicable and reliable method of forecasting seasonal and short-term streamflow, including floods, resulting from snowmelt and rain ;
3. Expand basic knowledge of the hydrodynamic and thermodynamic characteristics of snow through a program of fundamental scientific research ;
4. Increase knowledge of meteorological, climatological and hydrological phenomena as they influence 1, 2 and 3 above.

When these objectives have been achieved, they will be applied to limiting factors in the design of flood-control structures and of spillways of all types of dams ; to the improvement of flood-warning systems ; and to control allocation of industrial, municipal, power and irrigation water resources, i.e., reservoir regulation. Previous investigations have furnished very little knowledge about the water storage capacity, transmission rates and melting rates of the various types of snowpacks in the critical melting zone. Therefore, the study of the physics of snow is an important phase in our investigations.

Preliminary studies had been conducted at Soda Springs, California, by the Weather Bureau and University of Nevada, where the present Technical Supervisor of the Cooperative Snow Investigations, in conjunction with Dr. J. E. Church, studied existing recording precipitation gages and new seasonal storage gages under various conditions of exposure.

Additional cooperation is provided in a limited degree by three other federal agencies, Geological Survey, Bureau of Reclamation, and Forest Service, as they have a mutual interest in the objectives of this program. This cooperation is not in direct monetary contribution, but in assigning personnel to the Processing and Analysis Unit, assisting in stream gaging, processing streamflow records, mapping forest trail and trail bridge maintenance, etc.

ORGANIZATION

The **Program Director**, a Corps of Engineers employee, is in administrative charge of the program, and is responsible for the proper coordination of the various phases of the program. Also, he serves as a liaison officer between the several agencies and offices cooperating in this program. His office, together with the office of the Processing and Analysis Unit, is located in the South Pacific Division, Corps of Engineers, at the Oakland Army Base.

The **Processing and Analysis Unit**, a special organizational unit which bears the name of its functions : the processing and analysis of observational data. The work of the Unit may be classified as follows :

1. *Custody*, including a central depository for the recorder charts, fields notes, and other original data ; the data go through a standard process, including micro-filming for security and convenience, posting on control cards, binding and filing by location, element, station and period ;

2. *Appraisal*, referring to the observational data, and to the program as a whole; this procedure of continual self-examination provides flexibility for the program, whereby it can be modified promptly, as indicated by such things as :

- a. Increased knowledge of precision of instruments and observational techniques
- b. Increased knowledge of variation of elements to be sampled;
- c. Changing needs of the cooperating agencies; and
- d. Completion of specific projects.

3. *Transcription and preparation of summaries*, referring to data from recorder charts and other forms transcribed and summarized according to standard and rather objective procedures; the status of this phase of the work is shown on bar charts which accompany the monthly reports of operation; copies of the transcriptions and summaries are available, upon specific request to the Program Director, for use by participating agencies which need data in numerical form for study;

4. *Descriptive reports*, necessary for presenting the climatological and terrain environments in which the observations are taken; the three laboratories were deliberately located so as to study varying types of snow-streamflow regime, and, to apply the knowledge gained at these places, the interpolation must be based on known conditions;

5. *Analytical work*, following the extensional process of successive refinement of description, classification, comparison and correlation; the objectives of the Cooperative Snow Investigations have been expressed in the form of about fifty specific problems, which are periodically reviewed by the administrative and other offices of the Corps of Engineers and of the Weather Bureau for assignment of priorities; each problem is further subdivided into convenient work units known as projects, which are classified and listed in a technical report. Projects are allocated to individuals or teams in the Processing and Analysis Unit, and to other participants in the Cooperative Snow Investigations; for each project there is a file folder or set of folders for computations, graphs and text, each project being represented by a card in the control file; this control file presents a general view of the status of analytical work in optimum detail as it is examined and maintained, with provision for a detailed project statement to be kept in each folder, thus providing continuity of effort and reducing the possibility of wandering and losing the original objective; bibliographic work is an important function of the Processing and Analysis Unit, in whose files are several thousand citations of important scientific literature on snow and ice and related climatology.

6. *Technical Reports*, completed descriptive and analytical projects, disseminated in a limited distribution and serially numbered.

In order to accomplish the objectives outlined, it was decided that, initially, three snow research laboratories be established, selection to be made to cover three conditions of precipitation, i.e., (1) rain and snow with rain predominant, (2) rain and snow with snow predominant and (3) snow; with consideration given to other climatological and terrain factors in each laboratory area.

The **Upper Columbia Snow Laboratory** is a drainage basin of approximately twenty-one square miles in the Front Range of the Montana Rockies, at the extreme headwaters of the Columbia-Clark Fork-Flathead river system. It lies just west of Marias Pass on the Continental Divide, which forms its eastern boundary, and is partly inside Glacier National Park. It is traversed by the Great Northern railroad and U. S. Highway No. 2, which cross the Continental Divide at Marias Pass (elevation 5215 feet).

The laboratory area comprises three sub-basins, those of Skyland Creek, Upper Bear Creek, and Lower Bear and Autumn Creeks. Streamflow is so measured as to separate, from the rest of the area, Skyland basin, which has a deeper soil mantle and is more densely forested with lodgepole pine than the other basins. The land surface of Skyland basin has an average slope of approximately thirty percent and an average orientation toward the northwest.

The climate is characterized by a cold snowy winter (15° F. mean December to March), and a short, mild summer (55° F. mean July-August). During winter, fronts come across the mountains from the Pacific, or from the arctic plains of Canada to the northeast. Summers are generally continental in character, with considerable rain. The basin lies in the zone of sharp transition between the west-

coast climate with winter precipitation and the interior climate with summer rain; the total annual precipitation of approximately forty inches is fairly evenly distributed. Almost all the winter precipitation falls as snow, which accumulates to depths of about sixty inches (water equivalent of about twenty inches) near the end of March. Runoff averages about twenty-five inches, over half of which occurs during May and June.

Of the twenty-nine observation stations in the laboratory basin, nineteen are concentrated in Skyland basin. Headquarters station, at the confluence of Skyland and Bear Creeks, is near the center of the entire experimental area; in addition to facilities for servicing the field stations it has complete observational installations and living quarters for the laboratory staff.

The **Central Sierra Snow Laboratory** is a four-square-mile drainage basin just west of the crest of the Sierra Nevada in California, at the headwaters of the Yuba River, a tributary of the Sacramento. It is adjacent to the Southern Pacific railroad and U. S. Highway No. 40, which cross the Sierra divide a few miles east at Donner Pass (elevation 7190 feet).

The experimental area, Castle Creek Basin, has a moderately rough surface of granite and lava, bearing patches of soil and sparsely covered by sub-alpine forest, chiefly lodgepole pine; this surface is inclined toward the south-southwest at a mean elevation of 7500 feet.

The climate is characterized by winters that are cool and extremely snowy (temperature average, a few degrees below freezing) and by summers that are warm and rainless. Of the annual precipitation (major part snow, with occasional rain) of approximately sixty-five inches on the basin, seventy percent falls during the four months December to March. The snowpack begins to accumulate in November and, in spite of occasional depletion by rainstorms, increases to a depth of about one-hundred and twenty inches (water equivalent of approximately fifty inches) by the end of March. These phenomena result in the concentration of over half of the abundant runoff of water (approximately fifty inches per year) during two months in late spring.

The meteorologic and hydrologic processes are measured at twenty-four field stations distributed to sample the chief topographic factors, and at a well-instrumented headquarters station at the lower end of the basin.

The **Willamette Basin Snow Laboratory** is an eleven-square-mile area in the dense, wet forest of the middle Cascade range, some thirty miles east of the Willamette Valley and ten miles west of the high lava plateau of Central Oregon. It includes the headwaters of the Blue River, a tributary of the McKenzie, which flows west to join the Willamette near Eugene, Oregon.

Elevation within the experimental area ranges from 2000 to 5346 feet, and there are three major peaks: Squaw Mountain (5272 feet) at the north end; Carpenter Mountain (5346 feet) along the southeast side; and Wolf Rock (4503 feet), an isolated crag near the center. Almost the entire area is densely covered with heavy virgin forest of Douglas fir and hemlock-fir types. The surface has an average slope of thirty-five to forty percent, and an average orientation toward south and southwest.

The climate is maritime, with a seasonal shift in airflow from southwest in winter to northwest in summer. Mean air temperature varies from slightly below freezing during the winter to the high fifties in late summer. Precipitation is relatively heavy, totaling about one-hundred inches per year; maximum precipitation occurs in early winter, diminishing during late winter and spring to a minimum in late summer, which is nearly dry. It is established that at the four-thousand-foot zone half the winter precipitation falls as snow and that the mean snowpack amounts to about forty inches water equivalent by the end of March. Runoff from the basin is estimated to average about seventy-five inches annually, and is concentrated in the winter, the greatest portion occurring in the six months from November through April.

Living accommodations and headquarters have been constructed just outside the upper end of the basin at Westside Camp, which is reached from the South Santiam cross-mountain highway a few miles to the north. Installation of observational equipment began in the 1947-48 season with a water-stage recorder at the basin outlet and eight precipitation gages, of which seven are storage gages in the basin and one a recording gage at headquarters. It is planned to complete the field quarters and the installations to stress the study of hydrologic processes at the Willamette Laboratory in the summer of 1948.

The **Technical Supervisor**, by direct work with the individual laboratories, provides technical supervision of the laboratory program, and conducts or guides special studies in connection with the physics of snow. Technical reports covering the procedure, instruments and results of special investigations are prepared by him. Also, it is his responsibility to recommend modification of existing instruments, and to design and develop new instruments, procedures and experiments required for special investigations. Thus far, there have been developed :

1. Special instruments and experimental procedures for the measurement of radiation transmission through snow;

2. An electronic instrument for the measurement of the liquid water content of the snow in situ, and possible for determining transmission rate ;

3. Procedure and apparatus for the measurement of the liquid water holding capacity of snow;

4. Procedures for tracing water movement through the snow by means of dyes. Additional experimental procedures and instruments being developed include :

1. A special radiometer for measurement of long-wave radiation;

2. Electronic methods for measurement of water transmission rates;

3. Improvements in methods of electrical measurement of soil moisture;

4. Radioactive isotopes for use as possible melt and rain-water tracers, stream-flow indicators, and as activators for radio-telemetering systems to be used as snow and precipitation depth recorders;

5. Field tests of various types of precipitation gages;

6. Development of frost-proof housing for electrical contact type meteorological instruments;

7. Experiments on suitable anti-frost, anti-ice, and hydrophobic coatings for meteorological instruments.

THE PROGRAM AND TENTATIVE RESULTS

Hydrology, as a subject, must be viewed with proper perspective. Hydrology — unlike so-called exact science — is characterized by the fact that the factors comprising it are subject to very little control. Extraneous variables (while being permitted to vary systematically) cannot be eliminated in order to study those remaining. A drainage must be accepted as it is and the weather as it comes.

The **four broad objectives** of the Cooperative Snow Investigations have been expressed in the form of specific problems (originally thirty, now about fifty) outlined on five major sections :

Section I. Seasonal totals of elements :

A. Precipitation — Precipitation is measured at several stations in each laboratory;

B. Heat — The seasonal heat balance is being studied; some elements, such as incident and reflected short-wave radiation are measured directly, some elements can be estimated, and other elements, such as long-wave radiational exchange are not observed and can be estimated only with difficulty; preliminary estimates indicate that the net gain of heat by each basin during the melting season greatly exceeds the quantity necessary for melting the snow;

C. Streamflow — The seasonal total of streamflow is measured at each laboratory;

D. Hydrologic balance — Attempts to balance the water cycle in the laboratory drainage basins are still limited by the age-old equation of the hydrologic cycle, which has two unknowns — vapor exchange with the atmosphere, and variations in soil moisture and ground-water storage — it is believed that the yearly totals of these two unknowns each exceed the sampling and observational error in the precipitation and streamflow; also, it is believed that additional observations, with analytical techniques now available, will bring this problem to a successful conclusion.

Section II. Time distribution of processes :

A. Rates of precipitation — Precipitation mass curves are being prepared for stations at the laboratories; density of new snow has been found to be a function of air temperature, and additional study should define such a relation well enough to greatly enhance old records of snow depth; much analytical work remains in deve-

loping a method for indentifying the occurrence of precipitation as rain or snow by inferences or interpolations from regular observations at distant stations ;

B. Rates of heat transfer within the snow

1. Heat transmission coefficients of the snow have been determined, found to agree well with those of earlier investigations and to be highly correlated with readily observable physical characteristics of the snow; the albedo, or reflection factor, of the snow surface seems also to be correlated with easily observable physical properties of snow, but values lower than those usually reported have been observed, values less than fifty percent being common during melting of the snow;

2. Solar radiation has been correlated with regularly observed meteorological elements, providing a sound basis for estimating this element where pyrhelimeter observations are not made;

3. Studies on the depth of penetration of solar radiation into snow have been made with a special pyrhelimeter designed for the purpose; this instrument consists of a number of small copper plates, the upper surface of which is coated with a flat black, the lower surface to which a copper-constantan thermocouple is soldered and a small electrical heating unit cemented; the plates are supported in cavities in the snow and their temperature read by means of a deflection galvanometer in series with the thermocouple and a standard reference junction; by means of the electric heater unit the recorded temperature may be duplicated and the equivalent radiant heat computed;

Transmission coefficients for solar radiation have been determined with this instrument; for snow of thirty-five to forty-five percent density the measured transmission coefficient agrees with that obtained by other investigators; it was found, however, that the transmission coefficient is a function of snow density, less radiation being absorbed at the surface and transmitted through low-density snow than through high-density snow; the presence of considerable amounts of liquid water at or near the snow surface increases the amount of solar radiation absorbed without increasing the depth of penetration; preliminary results of this field of investigation are contained in a technical report :

4. Exchange of long-wave radiation has not yet been observed, although its importance is recognized, and some preliminary work has been done on design of an instrument for such measurement;

5. Heat transmission through the soil is a factor for which observations at a limited number of sites are available, but no analysis has as yet been made; freezing and thawing of snow and soil are being studied by means of several types of electrical resistance soil moisture units;

6. Heat transmission by turbulent exchange — both sensible and latent heat — is very important; while much work in this field has been reported from arctic investigations, the problem is greatly complicated by the natural factors of forest cover and rugged topography and little progress has been made to date;

C. Rates of melting — Melting rates are approximated by many periodic measurements of water equivalent or mass of snow, and by interpretation of stream-flow;

D. Rates of movement of liquid water — The distribution, storage, and transmission of liquid water in the snowpack are related to the structure of the pack;

E. Rates of runoff — The relation of runoff rate to other elements involves separation of contributions from rain and from melting snow; also there are involved such questions as the location of the melting zone and the time required for travel from various parts of the drainage basin; considerable progress has been made in appraising, improving and developing empirical methods of estimating daily runoff rates from melting snow; regressions on antecedent streamflow, observed air temperature, and other factors account for more than ninety-five percent of the variance in daily streamflow; it is believed that the methods developed will have useful application in forecasting or estimating the time of beginning of seasonal snowmelt and in separating the contributions from rain and melting snow; the shape of the diurnal hydrograph is highly correlated with other factors;

F. Rates of movement of water — Infiltration and transmission of water through soil are indicated by electrical resistance measurements; it has been concluded that much of the water from melting snow passes into the soil on its way to the stream;

G. Rates of evaporation and transpiration — Evaporation and transpiration are not observed directly, but analysis of other data indicates that during the winter

months the vapor loss from the snowpack has been so small as to be obscured by errors of sampling and observation; there are indications that most of the vapor loss occurs after melting starts, and is largely from shallow pools and warm soil moistened by the melting snow.

Section III.

A. Areal distribution of precipitation — Precipitation has been observed to vary greatly over the short distances of the laboratory basins, even in terms of total seasonal catch: minimum catch was as little as sixty percent of the maximum in the same basin, with no likelihood that either station represented extreme values; variations in precipitation over shorter periods was of course larger; the factors important to precipitation catch seem to be local exposure, orientation, concavity of the ground, slope and elevation;

B. Areal distribution of the snowpack and its characteristics —

1. Snow cover — aerial photographs at the Upper Columbia laboratory have suggested a recurrent pattern of melting year after year; analyses of snow cover against such terrain factors as orientation and steepness of slope have shown promise;

2. Water equivalent — the variation in water equivalent, or mass of the snowpack attributable to topographic factors is quite large; prior to the melting season, the station with maximum water equivalent observed in the basin averages twice as much as the station with the least; during the melting season this ratio rises to three or four times as much; among station means, the standard deviation is equivalent to a quarter of the basin mean before, and rises to about the same size as the basin mean during the melting season;

3. Density — variation in snow density was not excessive, usually encompassed by a range of ten percent;

4. Depth — depth of snowpack depends primarily on exposure and orientation and the range of variation (48 to 108 inches on 1 May 1946 in the Sierra Laboratory basin) indicates a dispersion similar to that of water equivalent;

C and D. Areal distribution of heat transfer and of snow melting zones — processes are extremely varied; efforts are being made to ascribe variations in radiation, temperature, wind and other elements, to effects of local exposure and of objectively defined terrain parameters;

E. Areal distribution of moisture in the snow — Storage, transmission and water-holding capacity of snow —

1. Calorimetric procedures have been used to determine the weight of unfrozen water in samples of snow from different depths within the pack during periods of high melting rates; the effect of storm rainfall on snow has been simulated by the addition of water to the surface of core samples of snow; the results of these studies indicate that where drainage is not impeded by impermeable ice planes within the pack or frozen soil the temporary water holding capacity of snow seldom exceeds fifteen percent by weight;

2. When the supply of water from melting or rain ceases, rapid drainage reduces the liquid water content of the snow to five percent or less within a few hours;

3. Temporary accumulation of liquid water in deep snowpack appears to be confined largely to the surface foot of snow and to a few inches at the bottom of the pack;

4. At densities between twenty-two and forty percent, little difference in the water holding capacity was found; in high density snow, ice planes or consolidated crusts are frequently present, and temporary impounding of water above such planes appears to be conducive to their rapid disintegration and increased permeability;

5. Observations of the movement of water through the snow to the soil indicate that horizontal travel of water through a pack may be as great as the vertical travel; also, it appears that the major portion of melt water, after percolating through the snowpack, filters into the soil and flows through the soil to a nearby natural drainage channel;

F. Areal distribution of Runoff — Areal distribution of runoff, on a small scale is probably related to areal distribution of the melting snow, but this is not being observed; variation in runoff among drainage basins is being studied by examination of standard records from many stations, in addition to those of the three laboratories.

Section IV. Generalization and transfer of conclusions involve two physical aspects :

A. Comparison of weather in the laboratory basins during these investigations with long-term records of the climate of these locations, and relations in the climatological environment; also, the weather occurring at the laboratories is described in terms comparing it with weather in regions where the results of the investigations are being and are to be applied;

B. The physical characteristics of the laboratory basins are described and analyzed so as to provide objective bases for transferring the results to drainage basins outside those being intensively studied.

Section V. Methods : This portion of the program involves studies of instruments, records, and techniques as guides in the technical prosecution of the program, control of accuracy and completeness of the observational program, and review of the adequacy for investigational purposes. There is involved the study of location, exposure, and spacing of instruments and the scheduling of observations.

SUMMARY

Considerable data have been accumulated and tentative results applicable to the basic objectives reached. There is still need to continue to obtain basic data and to develop other special studies pertinent to those already started on the physics of snow.



MAP OF WESTERN UNITED STATES
SHOWING LOCATION OF SNOW LABORATORIES
OF THE

WEATHER BUREAU—CORPS OF ENGINEERS COOPERATIVE SNOW INVESTIGATIONS

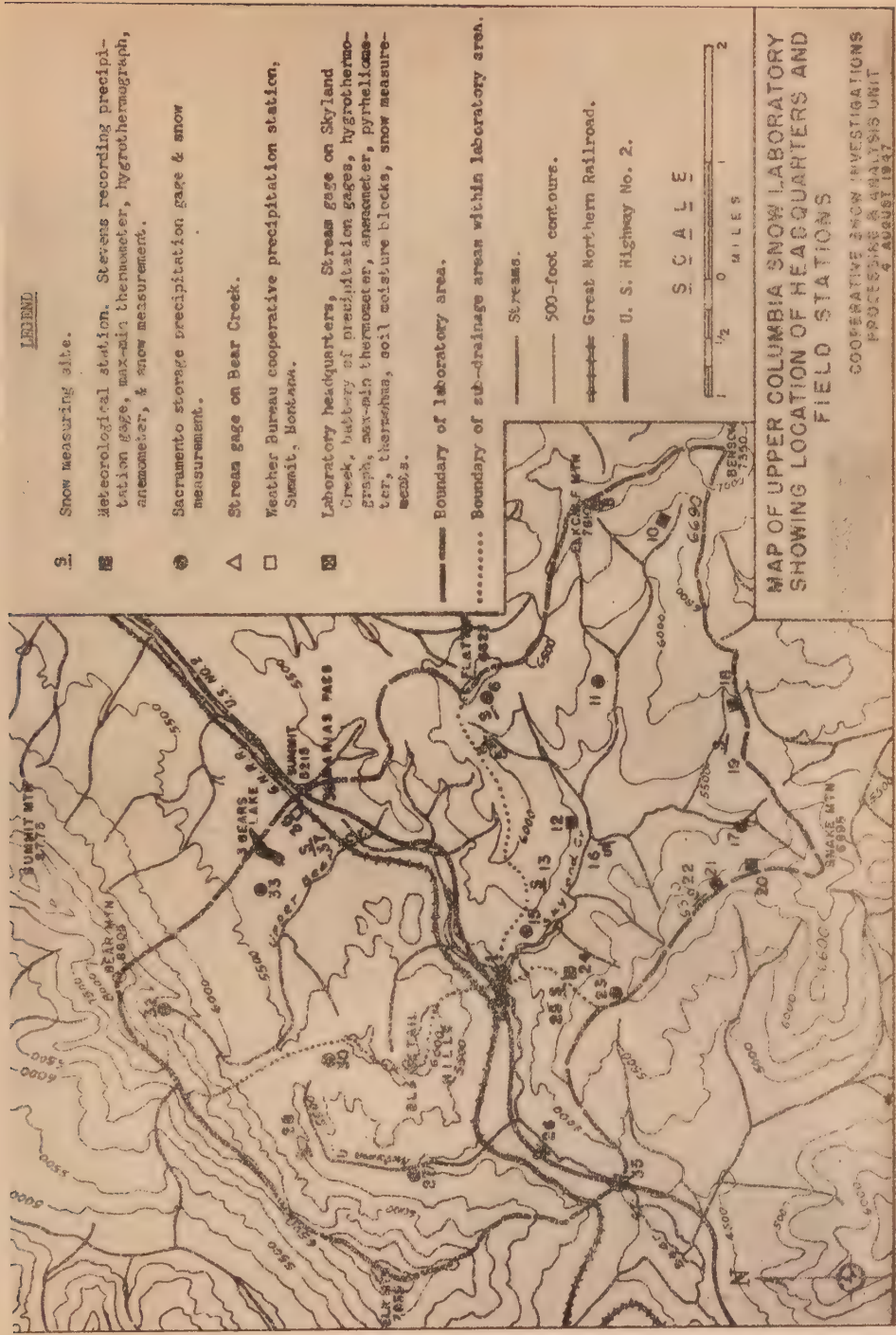




FIG. 1. Typical field meteorological station. This view of Station 5, Sierra Snow Laboratory, taken in December 1946, shows a Stevens record-precipitation gage and Alter type wind shield mounted on a 15-foot angle-iron tower. Instrument shelter houses maximum and minimum thermometers and hygrothermograph.



FIG. 2. Meteorological installation at laboratory headquarters. Instruments are mounted 4 feet above the snow surface, and 35 and 55 feet above ground to sample vertical variation in the elements. Picture taken in March 1946, Central Snow Laboratory.



FIG. 3. View of meteorological tower on top of ridge at laboratory headquarters. Shelters containing maximum and minimum thermometers, hygrothermographs, and thermohms are mountad 18 and 36 feet above ground. Wind vane, anemometer, and pyrheliometer are also mounted at top of tower. Picture taken on 24 April 1947, Upper Columbia Snow Lab.



FIG. 4. Pyrheliometer installation at laboratory headquarters. The pyrheliometer for incident radiation is mounted on top of tower. The pyrheliometer for reflected radiation is mounted under end of arm attached to side of tower. Each pyrheliometer is wired to a Micromax recorder Picture taken in September 1946, Central Sierra Snow laboratory.



FIG. 5. View of 200-inch Sacramento storage gage on 20-foot angle-iron tower at Willamette Snow Laboratory headquarters. Picture taken on 5 October 1947.

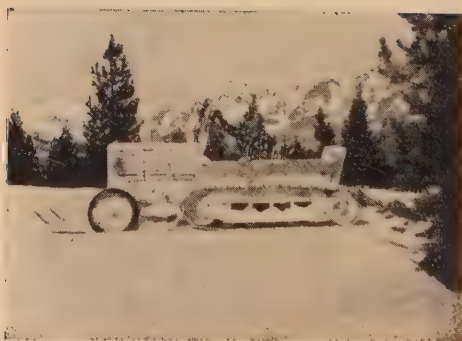


FIG. 6. View of tractor, snow (half-track), M-7. This type of snow vehicle is being used at all three snow-laboratories. It can operate as shown, with skis removed, or with wheels removed.



FIG. 7. View of 150-degree Vee-notch weir at end of 12-foot Parshall flume for measuring discharge in Castle Creek. For flows above approximately 75 cubic feet per second, the gate containing the Vee-notch weir is raised and flow is measured by Parshall flume. Picture taken on 17 October 1947 Central Sierra Snow Labor,



FIG. 8. Study of path of melt water through snow. Dark streaks on snow face show path of melt water from snow dyed with Fuchsine dye. Dye was applied to snow surface to right of gage stick. Picture taken at Central Sierra Snow Laboratory on 15 May 1947.

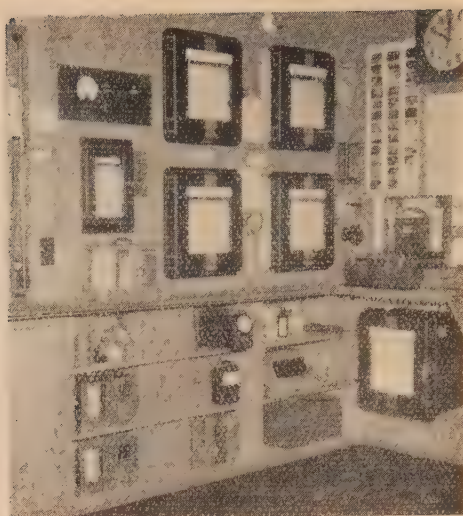


FIG. 9. View of instrument panel at Central Sierra Snow Laboratory. Shown are Mercurial barometer; microbarograph; Esterline-Angus recorder recording wind speed, wind direction, and sunshine duration; two Micromax recorders recording air, soil, snow, and water temperatures; two Micromax recorders recording incident and reflected radiation; and Buoyoucos bridge used for measuring soil moisture.



FIG. 10. View of thermohm trellis at Upper Columbia Snow Laboratory for measuring air and snow temperature at various levels. Thermohms and Buoyoucos blocks are also installed at various depths in the ground below the trellis for measuring soil temperature and moisture.



FIG. 11. View of coaxial thermocouple pyrliometer developed by Technical Supervisor for measurement of penetration of radiation through snow. Picture taken at Central Sierra Snow Laboratory in May 1947.



FIG. 12. View of capacitance meter developed by Technical Supervisor for measurement of free water in snow. Picture taken at Central Sierra Snow Laboratory in May 1947.

9) EFFECTS OF CHINOOK (FOEHN) WINDS ON SNOW COVER AND RUNOFF

by

O. H. HOOVER (Canada).

SUMMARY

With particular reference to Alberta and Western Canada, the characteristics and effects of Chinook winds are discussed. A concise statement is given on the Phenomenon and Maintenance of these winds. An exposition is cited showing how moisture-laden air at the Pacific Coast in winter may move easterly over the Continental Divide into a cold area and arrive as a dry air fully as warm as when leaving the coast.

An example of a Chinook which occurred in this area between March 17 and 24, 1939, is given and diagrams by meteorological and hydrological graphs illustrate the effect of this Chinook on snow cover and runoff. In the Appendix, meteorological data and daily weather maps, upon which the fronts have been plotted, show the initiation of this Chinook as the wind crossed the Rocky Mountains eastward.

Effects of Chinook winds on snow cover and runoff

The type of wind known as the Chinook in North America, which blows as a southwest or westerly wind on the eastern slope of the Rocky Mountains and eastward into the Province of Alberta, Canada, and into the States of Montana and Wyoming in the United States, is not only common to this area in North America, but occurs very probably from like causes at other points in various countries.

In Switzerland on the northern side of the Alps, a similar wind is known as the foehn, and blows from the south up, over and down from the Alpine range. In the south of Greenland on the west coast, during southeast winds, which have been elevated over the higher portions of the island, the temperatures, on occasion in winter, are reported to be distinctly higher than that of the atmosphere where the winds have originated.

Origin of name

Chinook Indians from the early days have inhabited the region of the Columbia river on the westerly side of the Rocky mountains in America, and owing to the supposition that these winds originated in this area, the name Chinook (probably first applied by Indians living on the east side of the divide) has from these early days characterized this unusual wind.

Characteristics and Effects of Chinook winds

Chinook winds in Alberta, and in particular at Calgary, always blow from the the southwest or west. Usually at the beginning the sky is overcast, excepting for the characteristic demarkation in the southwestern sky called the Chinook arch. This panorama comprises an arch-shaped strip of clear blue sky bounded underneath by mountain contours, and on top by a clearly marked edge of the clouded sky. As viewed from Calgary, the blue strip extends some 10 to 20 or more degrees above the mountain horizon, and if one travels toward the mountains it is found that the blue portion forming the arch frequently extends some 20 or 30 miles east of the most easterly mountain range, but this distance may vary greatly with various stages and intensities of Chinook winds. Through this arch-shaped strip the warm winds pour down the eastern mountain slopes and spread out over Alberta.

It is a commonplace experience that this Chinook-affected area in the dead of winter may be snapped out of a sub-zero condition into a balmy soothing state, at temperatures ranging from 25 to 50 degrees F or more above zero. On occasion in the city of Calgary it is possible, and has been experienced by the writer that, in walking westerly in the city, the temperature has changed from a sub-zero condition to that of at least 40° F above zero, in moving a distance of not more than a

city block. This marked the beginning of a Chinook. Not all Chinooks, however, originate in such a sudden fashion, but in most cases the change is much more gradual.

No precipitation appears to ever accompany these winds. The air is very dry and has a great affinity for moisture. To use a local expression it is said to lick up the snow, and in consequence, it uncovers the prairie range lands, making it possible in Southern Alberta for pasture grazing, during nearly every winter season. Irrigationists in this section often remark that the snow-fall during winter does not contribute very materially to the top soil moisture in the following spring. The main reason for this is due to the fact that much of the snow was lost to evaporation by Chinook winds.

Signs preceding a Chinook

Not in all instances does there appear to be a visible sign which will definitely forecast the approach of an on-coming Chinook. In very cold weather, however, these winds are often slightly preceded by a falling barometer, and the pressure remains relatively low while the winds last. Another indication which is believed to be a good sign at Calgary, (especially for the layman) is as follows : If the cold north wind veers gradually through the north, northeast, east, southeast and south, taking possibly 1 ½ or 2 days for the operation, generally the veering will continue to the southwest quarter, when a Chinook will develop. If this veering is rapid, a partial Chinook arch will form, but will not remain, and the wind will switch back to the north, with the consequent return of cold weather. Very seldom if ever, does the north wind back through the northwest to the west and southwest to produce these warm winds.

The phenomenon and Maintenance of Chinook Winds

The fundamental cause and related facts concerning the origin etc. of Chinook winds are not as yet believed to be fully understood. The inflowing Chinook air is generally Polar Pacific (symbol Pp). On occasion, however, it may be called Maritime Pacific (Mp) if warmer than usual, and originating farther south in the Pacific.

The advancing front of Chinook winds as revealed in aerological soundings have a distinct warm front up to great elevations. This front in moving eastwards leans forward over the cold mass at an angle with a slope varying from 1::100 to 1::300, and the former figure seems to be nearer correct when first crossing the Rockies. The cold air displaced by the Chinook and underlying the warm front is Polar Continental, or known in America as Polar Canadian, (Symbol Pc). The sudden change in temperature is owing to the rapid displacement of the Pc air.

In the atmosphere, winds are set in motion by the relative three-dimensional distribution of the temperature and pressure fields. The atmosphere really acts as a heat engine with sun as its fire, and in order to detect the action of this great prime mover in the local region of the Rocky Mountains, three major forces which act upon air masses may be considered.

1. The force due to the earth's rotation, called the Coriolis. With this may be considered the centrifugal force if the path of the air current is curved. As these forces are each necessarily at right angles to the motion, they do not involve the performance of work on the air particles, so are not the cause of the motion and will not be considered further.

2. The force of friction. This always acts parallel and opposite to the surface current. Its operation, however, deflects the wind obliquely across the isobars from *high* to *low*, and its equal and opposite reaction is found partly at right angles to the isobars in the form of a reduction of the effectiveness of the Coriolis force, and partly along the isobars in opposition to the gradient wind. Friction is work consuming, and hence is obviously not a cause of the motion.

3. The force due to the interaction of the temperature distribution and the pressure field, such as for example generates land and sea breezes. Since this force depends on action in three dimensions, its intensity can only be gauged indirectly from surface weather maps, by noting some phenomenon that involves work being done on particles of air. Such an indication is a zone with different barometric tendencies on the two sides. There is a component of wind known as the isallobaric

which blows from regions of rising pressure to regions of falling pressure, at right angles to the isallobars (lines of equal time rate of pressure change). It is independent of the instantaneous distribution of the isobars, and so may possibly blow from regions of *high* pressure to regions of *low*. In this way the pressure field would aid the wind motion, doing work upon the air particles and accelerating them.

In the case of Chinook winds, we know that they will occur when a *low* is deepening or moving in eastward, and forming up in northern Alberta. While barometric pressure falls in southern Alberta, it is building up on the coast; in other words, an isallobaric gradient is noticeable between the Pacific coast and southern Alberta. Therefore the Pacific air is being accelerated into the deepening *low*. In these circumstances it is usual to forecast a Chinook.

Here then is evidence of the atmospheric heat engine at work. The condition favourable for the local production of acceleration, a large gradient both of pressure and of temperature, is especially well fulfilled in winter in the Rocky Mountain region, owing to the juxtaposition of the cold, high-pressure Polar Continental air, and the warm, low-pressure Pacific air. The mountains may contribute to the suddenness with which the Chinook is accelerated by holding back the Pacific air until it spills through. This indication is manifest by a fairly steady and very large pressure gradient, with isobars parallel to the Rockies, which however, is greater than the observed winds would indicate. The additional heat in the Pacific air in the form of the latent heat of the moisture contained also plays its part, as will be seen.

It is known that the Chinook current of air, as it leaves the Pacific Coast on its eastward journey, is usually heavily laden with moisture, and the same air as it reaches Alberta is extremely dry. It is further known that this air, upon arrival in Alberta some 3000 feet or more above the Pacific Coast altitude, is as warm as, or warmer than when leaving the coast. The warm condition of this air occurs just after having traversed several high mountain ranges covered with snow, and without further consideration it would be reasonable to believe that a loss rather than a gain in heat would result.

This phenomenon, it is believed, has been correctly explained at various times heretofore, and without much doubt is the result of the following circumstances: When a mass of air having a relative humidity below the saturation point is lifted in the atmosphere, it expands and becomes colder at approximately the rate of 1°F per 180 feet, or 1°C per 100 meters raised. This is sometimes called the temperature lapse rate for dry air, the ordinary adiabatic lapse rate.

In the formation of a pound of water from water vapor by precipitation, some 970 B.t.us. or 244,000 calories of heat units are released to the air. Similarly in the freezing of a pound of water an additional 144 B.t.us., or 36,000 calories of heat are liberated. Hence in the creation of a pound of snow some 1,110 B.t.us. or 280,000 calories of heat are released. This is commonly called the heat of condensation. Some idea of the energy value for this heat transfer will be realized when it is remembered that 1 B.t.u. is the equivalent of 778 foot pounds of energy.

Pe air moving inland and eastward from the Pacific Coast toward British Columbia, is generally quite moist, and often approaching the dew point. This air, therefore, if caused to rise, will very soon be cooled below the saturation point, resulting in precipitation and consequent liberation of the heat of condensation, which in turn will retard the ordinary adiabatic lapse rate of the air temperature by about 50 percent.

This reasoning being correct, the following explanation for the appearance of warm and dry air in the Calgary winter Chinook winds, would appear to be in order. Moisture laden air at the Pacific Coast, in its movement eastward over several irregular mountain ranges, is mechanically forced upward to an altitude of possibility 10,000 feet. During the ascent the air cools at about 50 percent of the ordinary adiabatic rate, or at 1°F per 360 feet. With an initial temperature at sea level of 50°F , the temperature at 10,000 feet as the air passes over the Continental Divide, would be reduced to 22°F and a very large percentage of the moisture would have been withdrawn in the form of rain, hail or snow.

Descending the Rocky Mountains to Calgary's altitude of 3,400 feet, the air would be re-heated at the normal adiabatic rate of 1°F for 180 feet through a descent of 6,600 feet, or to a resultant temperature of 59°F . Some temperature reduction should be made owing to the fact that not all of the air is caused to rise to the 10,000 feet level, but comes through where passes exist at lower elevations. If 9° were allowed for this reduction, the air would then arrive at Calgary or points of like altitude at a temperature of 50°F .

Example

On March 17, 1939, one of these Chinooks began to develop in Calgary, and in two days' time about 10 inches of snow cover had almost disappeared. This warm wind continued until March 24, which is longer than the usual Chinook period, and maximum temperatures of over 70° F. were recorded at Lethbridge and many other points.

Plate 1 shows graphs of Relative Humidity, Temperatures and Sea Level Barometric Pressures at Calgary for a few days previous to and during the period of the Chinook of March 17 to 24, 1939.

Plate 2 shows by graphs the maximum and minimum temperatures for a few days before and during the period of the Chinook for representative stations from Victoria, B.C., at the west coast to Island Falls in Saskatchewan, about a thousand miles to the east. This temperature chart shows clearly the progress of the Chinook eastward by the pronounced temperature effect.

In this instance the winds penetrated deep into Saskatchewan, some 500 miles east of the Rocky Mountains, and small prairie streams which had been dry all winter were suddenly transformed into the maximum high run-off stages for the year. These streams normally develop spring freshet stages between April 10 and 25. On occasion, however, Chinook winds advance this spring flow period, and when such happens, as in 1939, the run-off is extremely rapid with maximum stages much higher than usual, but covering a shorter period of time.

Plates 3 and 4 show hydrographs of stream flow in cubic feet per second for 4 typical stream stations during March 1939. On plate 3 the flow for Ross creek at Irvine, 200 miles east of the mountains, is shown and also the run-off for Swift-current creek, some 120 miles further east. Plate 4 shows similar flow data for Lodge creek and the Frenchman river located on the 49th parallel of north latitude at 160 and 270 mile intervals respectively east of the Rocky Mountains. These demonstrate very clearly the effect of the warm Chinook winds on a snow-covered prairie section of country and they also show that in a matter of 3 or 4 days the streams of such an area may assume peak run-off stages for the season.

In the Appendix, Table 1 shows meteorological data for the 7 day period March 18 to 24, 1939, as obtained by the United States and Canadian Meteorological Services. From this data it will be observed that the air having a mean temperature on the outer coast and lower mainland varying from 47.5° to 49° F., on arrival at prairie points within 200 miles east of the mountains, has a mean temperature of about 49° F., and some of these prairie towns has temperatures of 20° to 30° F., below zero only 2 or 3 days prior to the beginning of the Chinook.

It will be noted that during the Chinook winds' period very heavy rains occurred at certain outer points along the British Columbia Coast. At Rivers Inlet for instance, 8.83 inches of rain was recorded in 7 days. At inland and mountain points in British Columbia some rain was recorded but in a more spasmodic manner. On the eastern side of the divide, however, where the Chinook had become effective, no precipitation was recorded.

Plates 5 to 9 show daily weather conditions for Canada from March 17 to 22, as prepared from the daily Weather maps as issued by the Meteorological Division, Canadian Department of Transport, Air Services. Air fronts have been drawn on these maps showing the advance of the March, 1939, Chinook.

Acknowledgment

Able assistance was given the writer by the Controller and Officials, particularly Mr. F. Graham Millar, of the Meteorological Service of Canada, in the collection of information and data contained in this report.

Graphs Showing Relative Humidity, Temperatures Fahrenheit, and Sea Level Pressures at Calgary

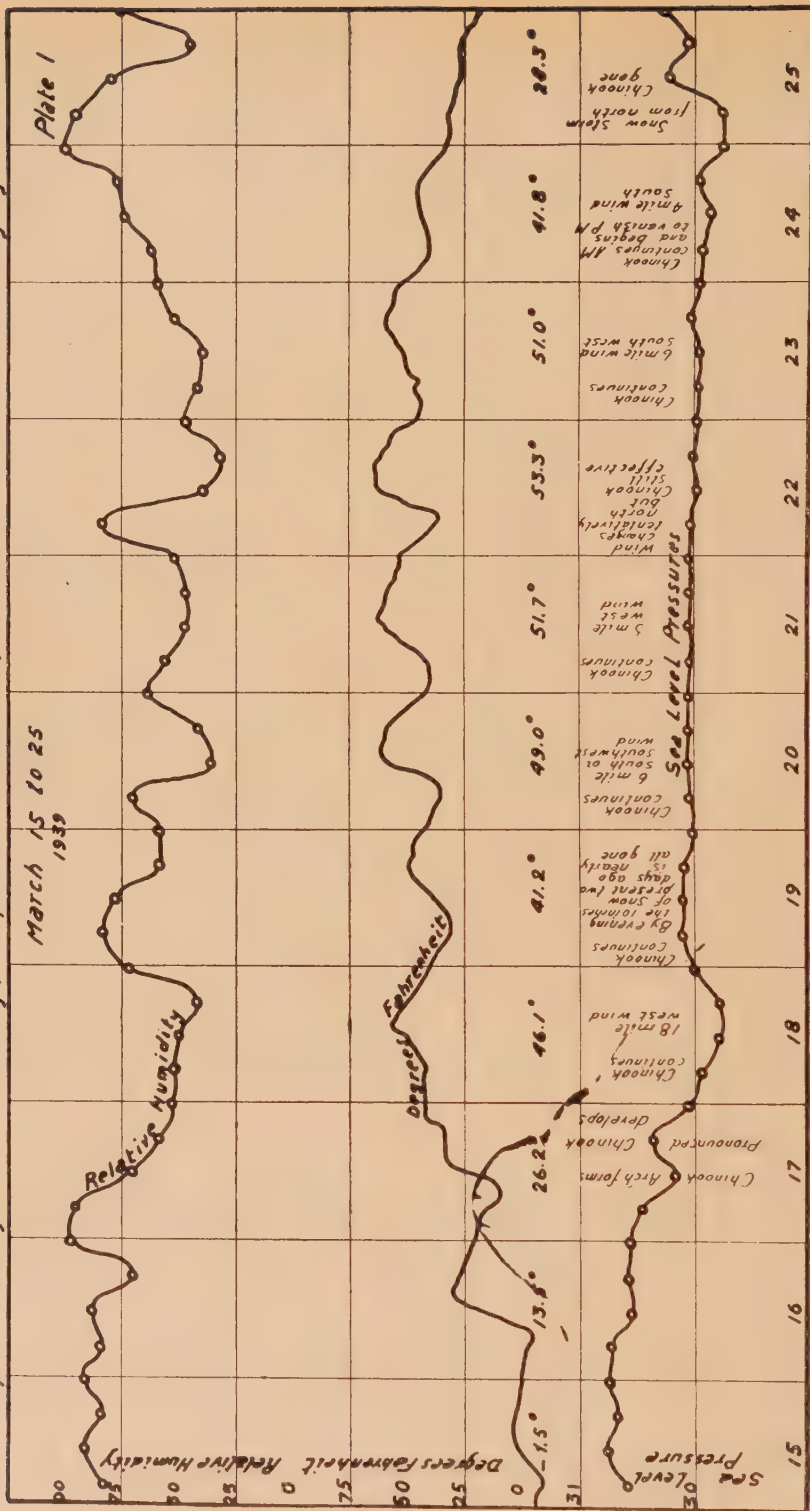


Fig. 1

Graphs Showing Maximum and Minimum Temperatures in Degrees Fahrenheit for several Stations in the Path of the Chinook of March 17 to 24 1939

Plate 2

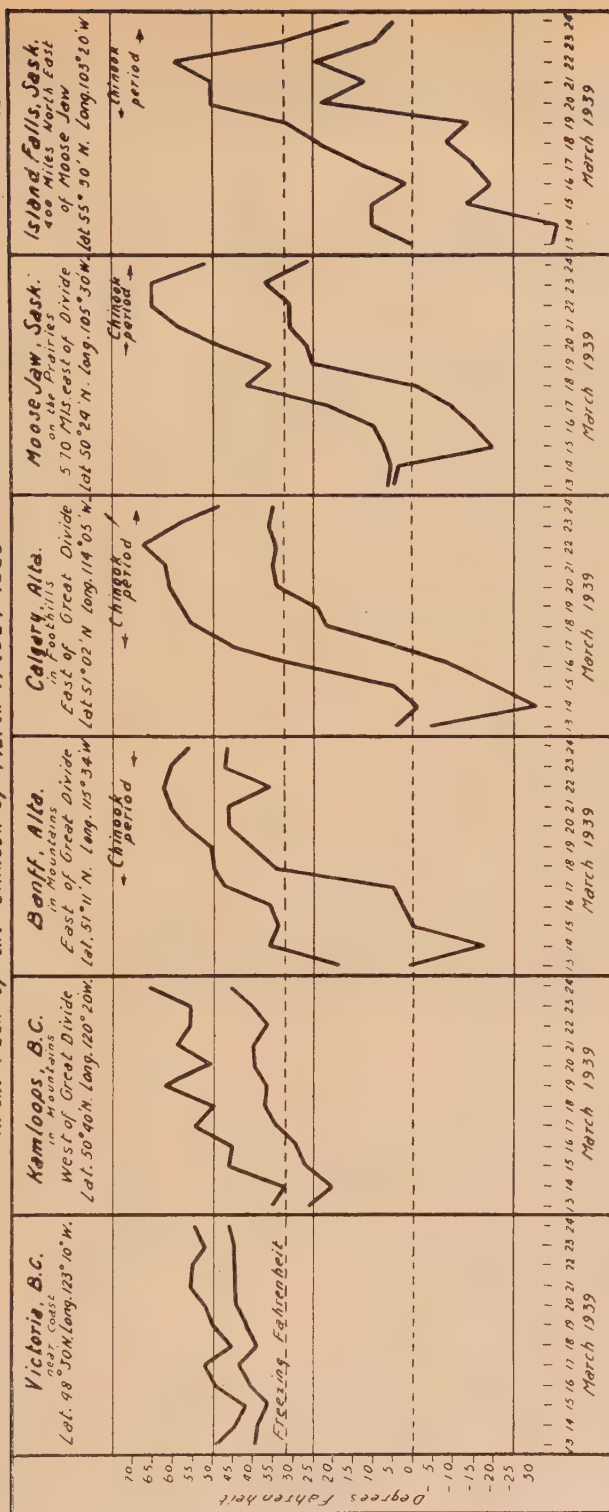


Fig. 2

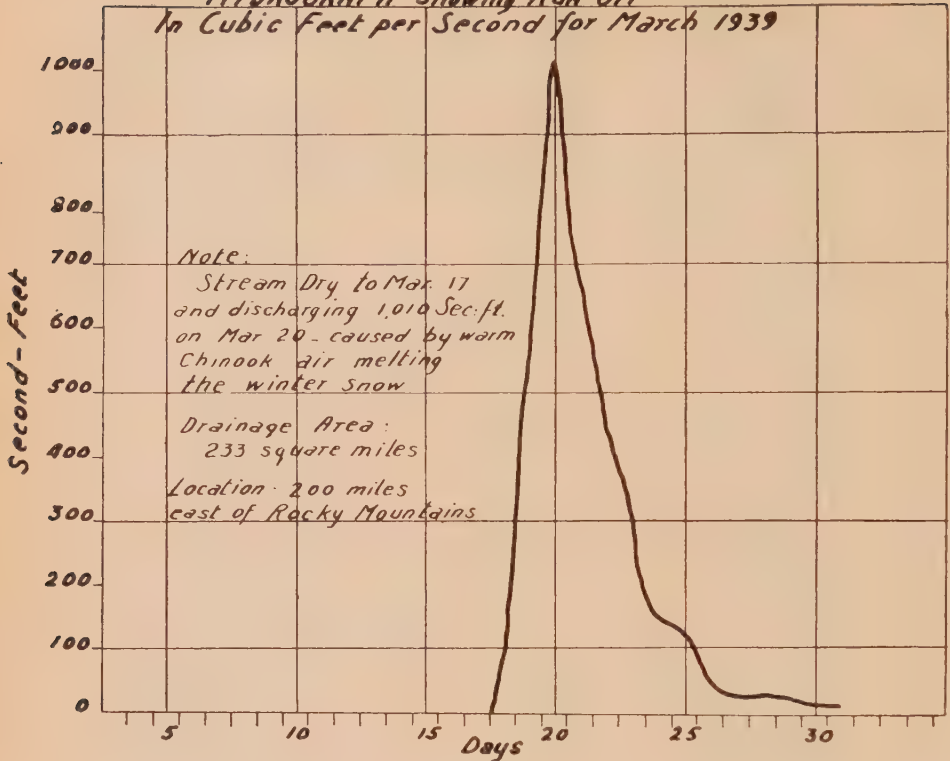
Ross Creek at Irvine

Lat $49^{\circ} 37' N.$; Long. $110^{\circ} 16' W.$

Plate 3

HYDROGRAPH showing RUN-OFF

In Cubic Feet per Second for March 1939



Swiftcurrent Creek at Swiftcurrent

Lat $50^{\circ} 17' N.$; Long. $107^{\circ} 48' W.$

HYDROGRAPH showing RUN-OFF

In Cubic Feet per Second for March 1939

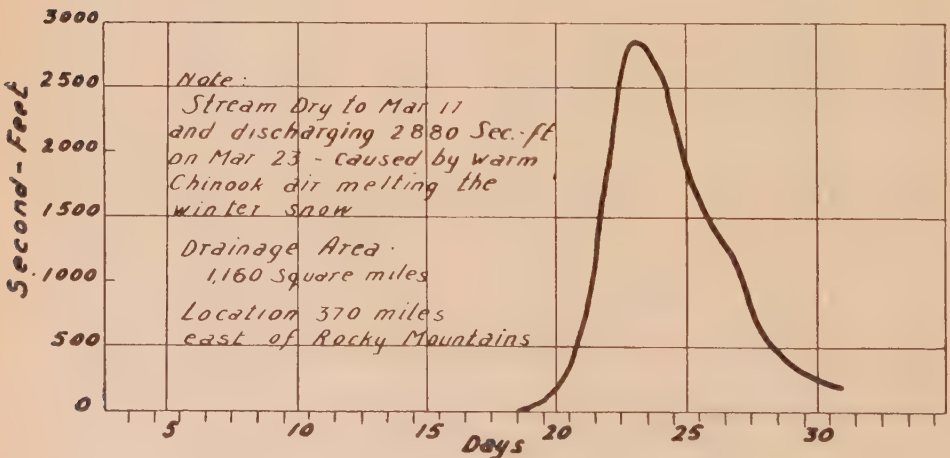
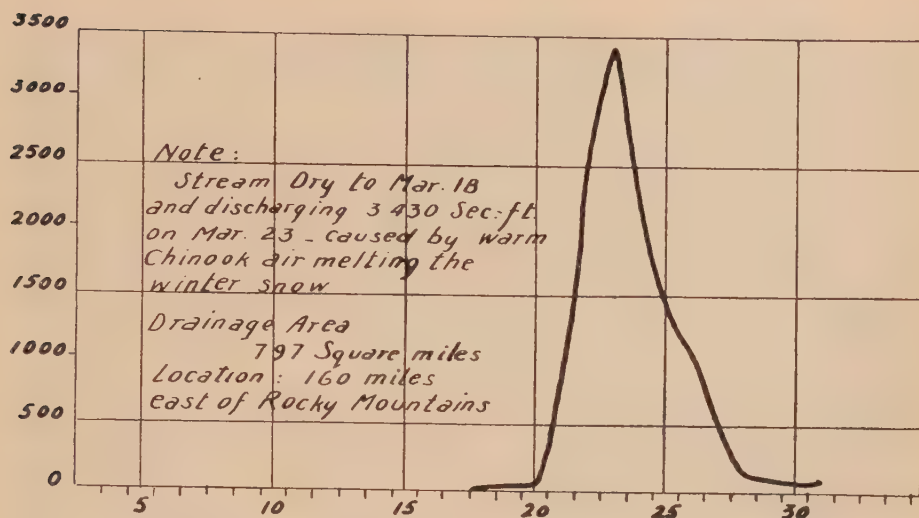


Fig. 3

Lodge Creek at International Boundary

Lat. $49^{\circ}00'$ N.; Long. $109^{\circ}45'$ W.

HYDROGRAPH showing RUN-OFF Plate 4
In Cubic Feet per Second for March 1939



Frenchman River at International Boundary

Lat. $49^{\circ}00'$ N., Long. $107^{\circ}17'$ W.

HYDROGRAPH showing RUN-OFF
In Cubic Feet per Second for March 1939

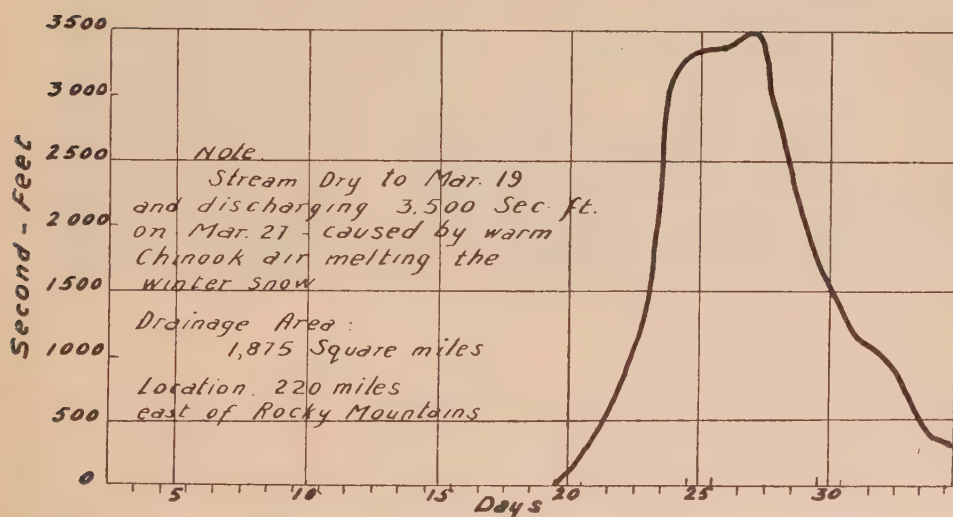
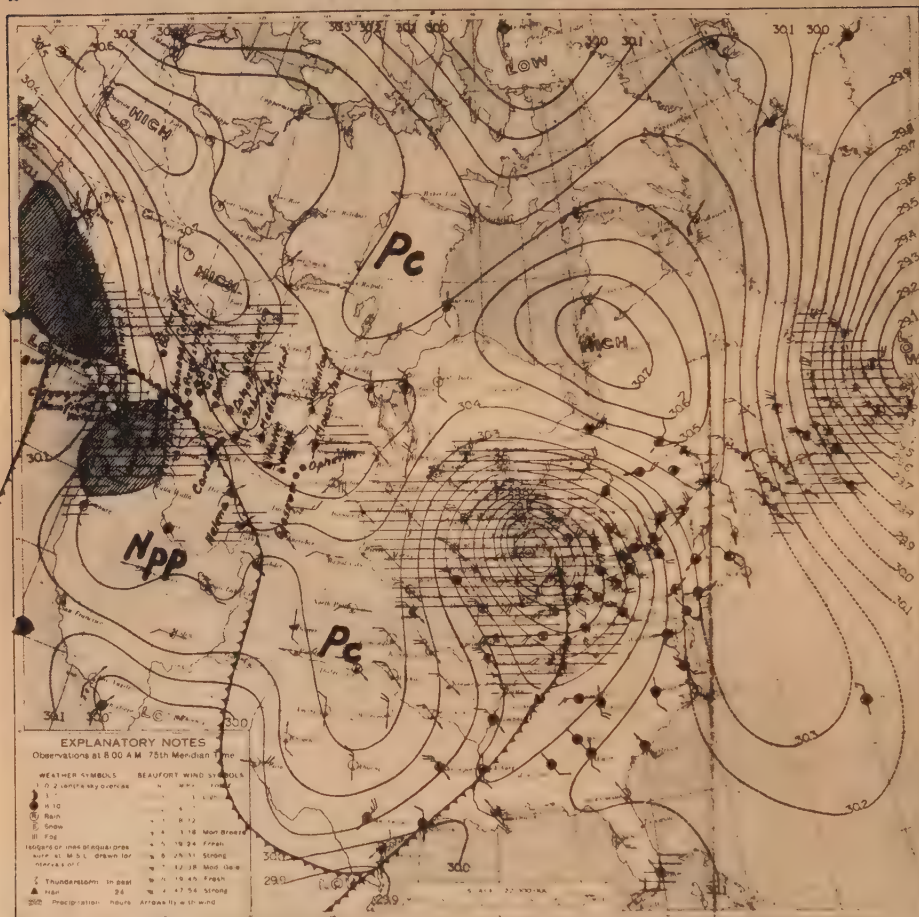


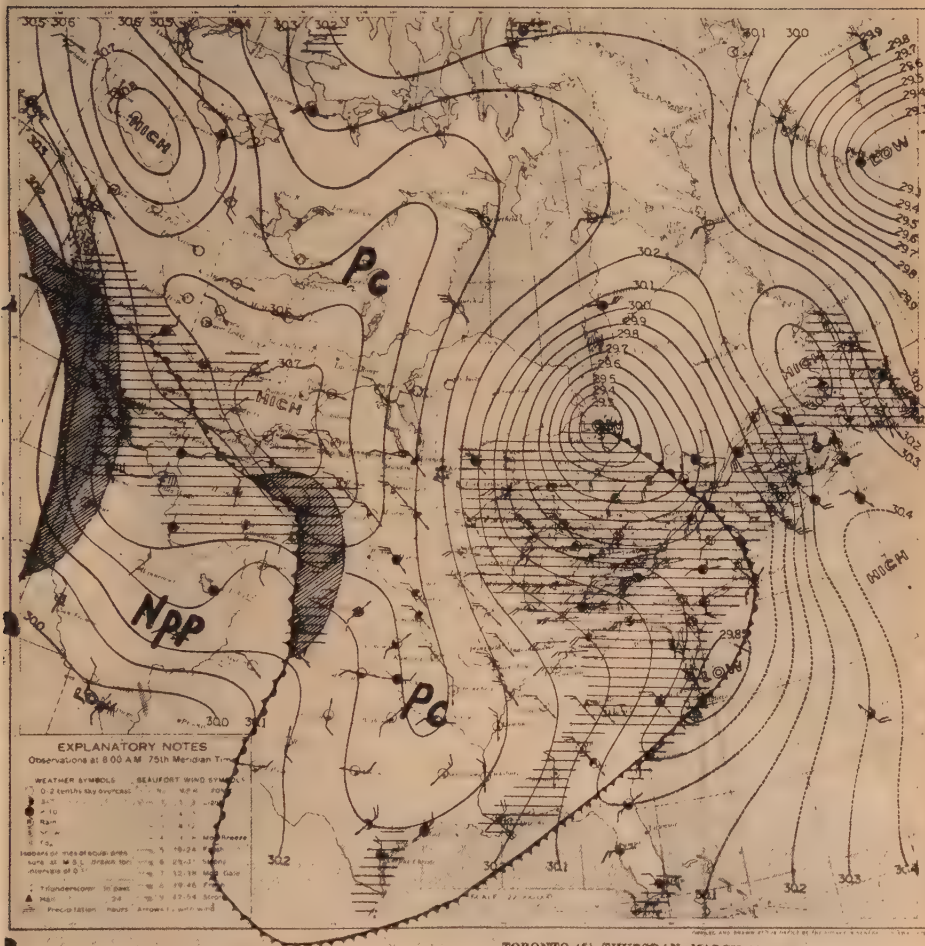
Fig. 4

CANADA—DEPARTMENT OF TRANSPORT
 Mar. 15, 1939 AIR SERVICES—METEOROLOGICAL DIVISION *Plate 5*
DAILY WEATHER MAP



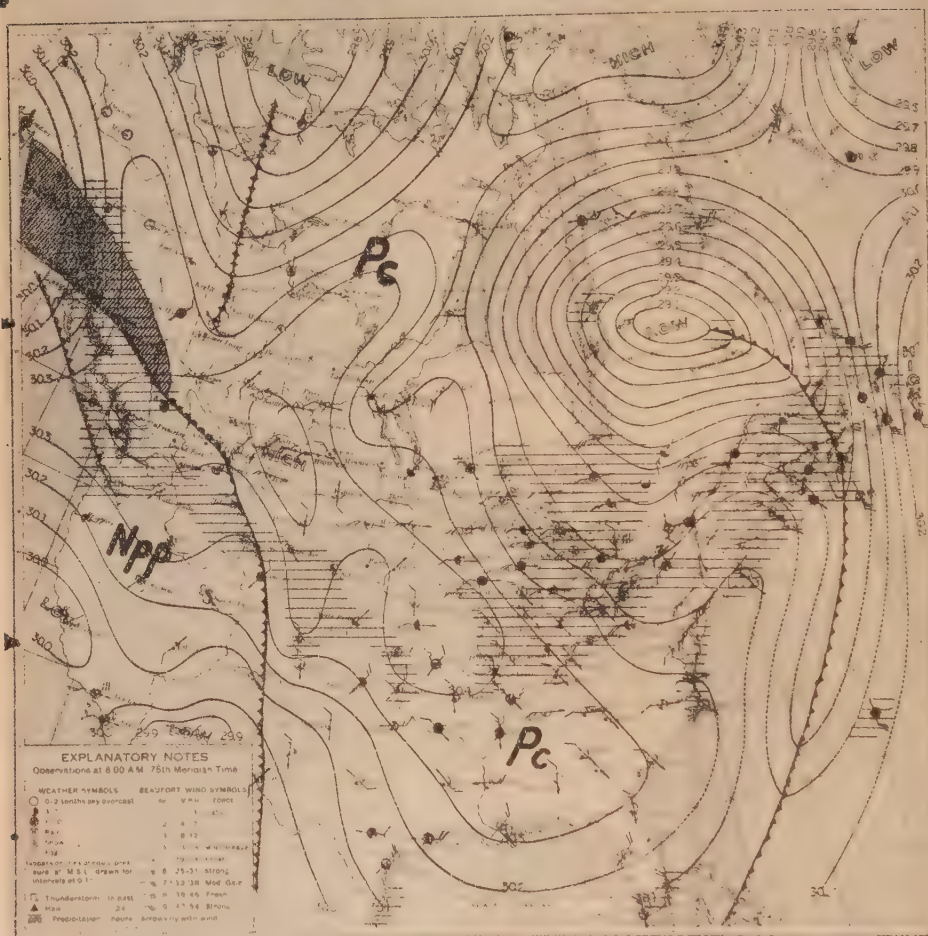
TORONTO (5) WEDNESDAY, MARCH 15, 1939

DAILY WEATHER MAP



Mar. 17, 1939

CANADA—DEPARTMENT OF TRANSPORT
AIR SERVICES—METEOROLOGICAL DIVISION *Plate 7*
DAILY WEATHER MAP



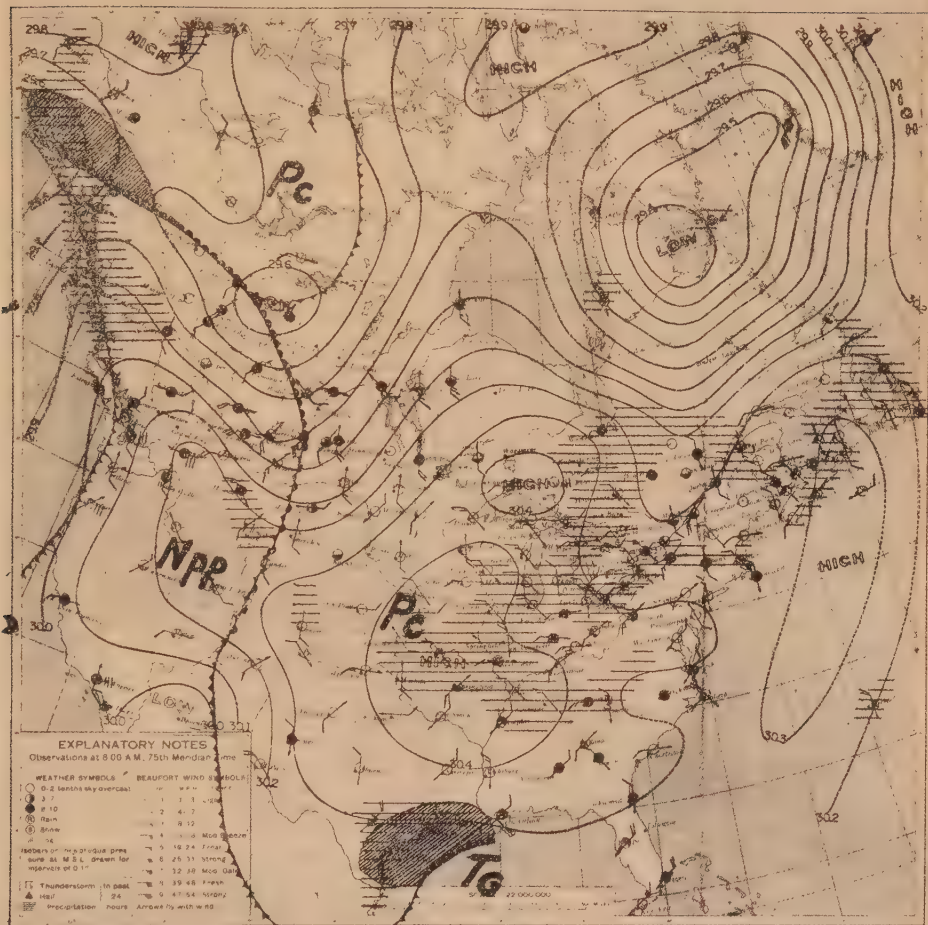
TORONTO (5) FRIDAY, MARCH 17, 1939

Mar. 18, 1939

CANADA—DEPARTMENT OF TRANSPORT
AIR SERVICES—METEOROLOGICAL DIVISION

Plate 8

DAILY WEATHER MAP



TORONTO (5) SATURDAY, MARCH 18, 1939

CANADA—DEPARTMENT OF TRANSPORT
AIR SERVICES—METEOROLOGICAL DIVISION *Plate 9*
DAILY WEATHER MAP



APPENDIX

TABLE I
 Meteorological Data March 18-24, 1939.

Outer British Columbia Coast

Station	Mean Temp. °F	Highest Temp. °F	Lowest Temp. °F	Precipi- tation Ins.	Wind
Rivers Inlet.	—	—	—	8.83	—
Quatsino.	47.0	54.0	37.0	1.78	—
Clayoquot	48.0	58.0	41.0	2.40	S. E. & W.
Bamfield.	—	—	—	1.65	SW; NW; W & S.
Mean	47.5° F				

Vancouver Island.

Victoria	47.4	56.4	38.4	0.42	N; W, Calm S. W.
----------	------	------	------	------	------------------

Lower Mainland.

New Westmintser	50.6	73.0	37.0	1.20	—
Agassiz	52.4	72.0	34.0	1.28	—
Kamloops	47.7	66.5	37.0	—	variable
Okanagan Centre.	45.6	67.0	32.0	—	—
Penticton	49.1	70.0	38.0	—	variable
Mean	49.1° F				

Mountains

Revelstoke	37.7	61.0	26.0	1.60	—
Glacier	34.9	48.0	21.0	0.63	—
Golden	41.6	59.0	24.0	0	variable
Barkerville	39.6	52.0	8.0	0.21	S. S.W.
Banff	49.3	62.0	35.1	—	S.W.
Helena	49.4	67.0	30.0	—	—
Mean	42.1° F				

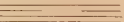
Prairies (within 200 miles of Mountains)

Station	Mean Temp. F.	Highest Temp. F	Lowest Temp. F	Preci- pitation Ins.	Wind
Cardston.	50.4	68.0	22.0	—	—
Lethbridge	53.3	73.9	30.0	—	SW.
Calgary	48.8	67.3	22.0	—	W., S. S. W.
Medicine Hat	47.4	71.0	24.5	—	S. W.
Havre	51.6	76.0	17.0	—	—
Mean.	49.3° F				

Prairies (more than 200 miles east of mountains)

Malta, Montana	42.9	71.0	3.0	0	—
Glasgow, "	46.0	74.0	6.0	0	—
Opheim, "	40.0	61.0	10.0	0	—
Edmonton, Alta	38.2	59.0	21.0	0.21	S.W.
Moose Jaw, Sask	38.6	65.0	0	0	—
Saskatoon, "	32.8	44.0	21.0	0	—
Mean	39.8° F				

METEOROLOGICAL SYMBOLS USED

Isobars	
Surface cold front	
Fading cold front	
Surface warm front	
Surface occluded front	
Cold frontogenesis	
Stationary front	
Polar Continental Air	Pc
Polar Pacific Air	Pp
Modified Polar Pacific Air	Npp
Tropical Gulf Air	Tg
Rain, or precipitation area	

10) ON THE RELATION OF PRECIPITATION
 ACCUMULATION AND MELTING OF SNOW
TO THE STREAM FLOW IN THE SAN JUAN RIVER,
 ARGENTINE

by

G. HEINSHEIMER

It is not the purpose of this short study to offer a complete solution of a very complicated problem. What is pretended by the relater is rather to communicate to the experts some points of view, hoping that their observations and opinions, objections and critic on the matter may be uttered or notified to the Hydrogeology and Geology Department, Reconstrucción de San Juan, Ciudad de San Juan, Rep. Argentina.

It is surprising that during the last winter (1946), more snow fell in the Cordillera than in the year before. Notwithstanding the volume of water was very moderate in the San Juan river, as well as in its two components, the Río de los Patos (S) with de Río Blanco and the Río Castaño. These rivers are the only drainage of about 25.000 km² of the Cordillera de San Juan and moreover form the vital artery of the Province where agriculture and vine growing depend exclusively on artificial watering. What is the motive of such an unfavorable behaviour ?

1. In the first line it is evident that the whole water content of the river originates from the atmosphere, other water supplies being negligible. Hence the field of investigation narrows considerably, since the Precordillera and the surroundings of the capital city, which also the river runs through, belong to the dry zone with less than 100 mm of annual rainfall. The 25 year normals of the 4 San Juan-Province-stations of observation amount during 1913 to 1937 from 83,8 to 99,0 mm, (1) therefore this part of the river basin must be considered as a region of losses by evaporation and infiltration, while for alimentation rests only the Cordillera.

As to the quantities of precipitation in this region I could not find more than a single notice. Carlos Wauters (2) writes : « Hasta un solo dato aislado nos ha proporcionado la primera estación de altura del Cristo Redentor en su año inicial, el observador destacado allí calcula que en el último invierno la caída de nieve fué superior a los 6 m ». Until now the first (meteorological) high station Cristo Redentor (2832m over sea level) has provided us with one isolated observation during its year of beginning; the observer calculates the snow fall during the last winter (1939 ?) of more than 6 m (3).

But even that information is very incomplete, since it is a well known fact that in the mountains the precipitation varies considerably even in places of the same altitude and very small distance; moreover the water equivalent of the snow may oscillate from 0.004 to 0.65 of the same volume of water.

To collocate totalisators in the Cordillera is a very expensive task, and even more so their maintaining and regular observation. In spite of it their value is very questionable as long as there is not a dense net of them available. Therefore the relater thinks it to be more advantageous to follow the plan exposed by A. de Quervain (4) and collocate permanent levels cemented to rocks or floating ones in the glaciers — the more of them the better — which must be read at least once a year, if possible during spring and summer, before the snow loss of autumn occurs. Furthermore the water equivalent of the snow has to be determined and the surface marked by a big spot of yellow or red ochre. Of course there must be given consideration to the fact the snow will accumulate in the valleys at costs of the adjacent hills and slopes.

2. But besides the precipitation also the *conditions of the snow melting* are of greatest consequence to the stream flow of the rivers. I may presume that the work of J. E. Church and his collaborators is well known to the reader. According to their experience in the south of the U.S.A., the water volume prediction bases itself

on the water value of the snow sheet existing on April 1st; the runoff is moreover considerably influenced by the precipitation during the period of melting, that is in April and May. On the other hand precipitation during the summer and autumn, before the snow is gathering are not of great importance to the stream flow of the rivers, motivating at the best increases of short duration.

Converted to the conditions of the Cordillera de San Juan the term should be fixed on October 1st or even later, referring to the great height of the mountains. In the literature read by me I found a reference which seems to coincide with such a supposition. Carlos D. Storni (5) writes : « Por otra parte conviene recordar que la fuente permanente de la riqueza de San Juan está constituida por la irrigación que alimenta el Río del mismo nombre y su candal depende de los deshielos en los glaciares, fenómeno vinculado con la temperatura reinante al final des equinoccio de primavera, la cual tiene influencia preponderante. En efecto, saben los hombres prácticos de Calingaste, San Juan, que caundo a fines de Noviembre y principios de Diciembre el tiempo transcurre fresco, no baja agua en el Río de las Patos, porque no hay temperature que provoque el derretimiento intenso de la nieve. En estas condiciones las crecidas de los rios se postergan, hay extensas porciones de cauces sin aguas y falta asi una elevada proporción de la infiltración ». On the other hand, there must be remembered that the wealth of San Juan depends on the watering nourished by the river of the same name and the volume (of water disposable to that purpose) relies on the melting in the glaciers, phenomenon vinculed with the temperature prevailing on the end of the spring equinox which is of the greatest importance. Indeed, the practical man of Calingasta (village in the big tectonic furrow between the Cordillera and Precordillera) do know, that no water is coming down in the Río de Los Patos (that seems apparently : « small quantities of water », for this river never dries up completely) when the time of the end of November and beginning of December is cool; because in this case there is no temperature which could incite an intense snow melting. Under such conditions the increase of water in the rivers is belated, big stretches of the river beds remain dry and a considerable quantity of infiltration fails to happen (3).

The last sentence is somewhat surprising, because dry stretches in the river beds can't occur unless the water has disappeared in the higher parts by seepage. /Of course abstraction made from evaporation, which will dry up very small streamlets only; and also from retention in the soil or in reservoirs like lakes etc. /. I am not able to find out, what Storni thinks unless it would be, that penetration to the water table might be higher, if also these temporally dry stretches would contribute.

There is no trial made in order to explain the phenomenon, notwithstanding that such an explanation doesn't seem very difficult. In general, all the valleys of the Cordillera and particularly the great tectonic furrow between the Cordillera and the Precordillera are replenished by enormous masses of debris of so gigantic dimensions, that they are able to swallow any quantity of water which may come down. Only in the case when the interstices inside the superior sheets of the debris are closed, for instance by ice, and when the melting proceeds early in the spring and rapidly, so that in spite of the hot sun of these latitudes, the fusion water doesn't become heated enough for thawing the ice; only under conditions of that kind the water of some tributary torrents will reach the main river. Hence to my opinion just the contrary to Storni's presumption is correct; when stretches of the river beds stream with water, while they are dry in normal times, then nothing or very small quantities of liquid will penetrate into the underground. The alternative is : superficial or underground water. Both can't happen simultaneously, but on very rare occasions, when extraordinary masses of precipitation meet with an early and rapid fusion in the valleys which continues later on slowly up to the heights.

A very good example provides the drain of the Cordillera de Ansilta. All the torrents produced by this part of the Cordillera de San Juan, which ascends to nearly six thousand metres of elevation, dry up superficially in normal years, when entering the main valley. But I have been informed that in 1941, when the melting of the snow began early in the spring, they continued until reaching the Río de Los Patos (6). That certainly must happen sometimes, as it is showed by the deep and broad dry beds, incised in the enormously large main valley for many miles, from the opening of the side valleys down to the deepest furrow where the Río de Los Patos flows.

The zone of alimentation of the San Juan river with its 25.000 km² in the Cordillera is vast enough that 100 mm of annual precipitation would produce a water volume of 80 m³/sec., if this precipitation would reach the river without losses. But the real amount of precipitation is certainly by far bigger; we remember the

observation before mentioned in Cristo Redentor, which corresponds to at least 600 mts. of water during the winter only.

Since the average annual water volume of the river amounts to $61 \text{ m}^3/\text{sec.}$, a) it may be estimated that $7/8$ more or less of the water provided by precipitations get lost by evaporation and infiltration.

b) the time when melting in the Cordillera begins, and the velocity of its progress seems to be of more importance to the stream flow than even the amount of precipitation.

c) the great elevation of the Cordillera always tends to delaying partly and distributing the fusion, hence aids towards the continuity of the water volume in the river.

d) The glaciers are small, situated at great altitudes, exposed to the evaporation and radiation of a very dry climate, hence very cold; therefore they supply less water than for instance Alpine or Rocky Mountains glaciers of the same size would do. Moreover they cover relatively smaller part of the area of the river Basin. Hence they are not as efficient regulators and compensators as Alpine glaciers, which yield water during dry and hot summers and store it during humid and cool periods.

3. Stream flow in the San Juan river (7).

a) Distribution of maximum and minimum stream flow — table 3 — The average and minimum stream flow of 37 years (1910 to 1946) amounts to Ma. — 282 and Mi. — $30 \text{ m}^3/\text{sec.}$ respectively and is distributed over the year as shown by table 3.

Maxima occur during 4 months only: November, December, January and February. On the other hand the minima are distributed the whole year round with exception of March. Hence March didn't contain any extreme runoff for 37 years.

From the frequencies (columns 2 and 7) follows that the *normal water year* culminates in a *January maximum* stream flow which averages nearly to the amount of the general average of 37 years maxima (Ma. — $4 \text{ m}^3/\text{sec.}$ or $1 \frac{1}{2} \%$). The corresponding *minimum* occurs in *August or September* and also nearly coincides with the general minimum, Mi — $1 \text{ m}^3/\text{sec.}$ or 3% .

When highest water volume appears in *December*, its average is much higher: $419 \text{ m}^3/\text{sec.}$ — Ma. + $136 \text{ m}^3/\text{sec.}$ or 49% , but that is only due to floods. Indeed 5 out of 10 cases stayed below average.

November maxima, motivated by premature snow melting of a relatively small stock of accumulated solid precipitation — otherwise the highest water would appear later in the year — are *always low*, only in one out of eight cases not very far above average.

The same occurs yearly with February maxima, although the table doesn't show it. That is owing to the fact that the highest February maximum of $626 \text{ m}^3/\text{sec}$ occurred on February 1st. 1920. If we take it to January, it will increase the January maximum with 23 at $301 \text{ m}^3/\text{sec}$ or $+ 8 \%$ approximately; while the February maximum average would get down from 280 to $103 \text{ m}^3/\text{sec.}$ or Ma — $179 \text{ m}^3/\text{sec}$ or $- 63 \%$.

This case like that of the December maxima shows, how cautiously statistics of runoff must be used. The errors to which we are led here, are due to the fact that the minimum stream flow oscillates in 37 years up to six times its lowest rate; but maxima stream flow up to 24 times its own lowest value and even 78 times the lowest minimum stream flow of the period. Hence high water has a much lower relative weight, which may lead to odd conclusions if not duly considered.

In table 4 I added to each annual maximum the next following minimum (8). It shows the highest runoff is likely to occur when the maximum appears in December, the minima in this case being over average also.

To small maxima due to premature melting (November) or belated melting (February) belong low minima and therefore probably also a small total runoff.

From different publications of a more historical character (9) we find catastrophic inundations arose nearly every year in the first third of the nineteenth century, culminating in the disaster of December 1833 and January 1834. Are they in connection with the glacier high stand happening in the Alps at the same time?

There are no estimations available as to the volume of water of these floods. But that of Christmas 1888 and New Year 1889 was estimated to $3.000 \text{ m}^3/\text{sec}$; that of 1906 to $2.000 \text{ m}^3/\text{sec}$.

The dates of the floods as far as I could find out underline the fact that highest water bearings are likely to produce themselves in December or January.

The relation of the highest water known to us (3.000 m³/sec.) to the 37 years average maximum is 11 : 1, to 37 years average minimum 100 : 1, to 37 years absolute minimum 214 : 1.

b) Average monthly and annual runoff.

The values of 12 years, 1935 to 1946, are shown in table 2.

The 12 years annual average A amounts to 60, 8 m³/sec.

The annual average during this time oscillates from 31,4 m³/sec. (A — 28,4) in 1939 to 122,2 (A + 61,4) in 1942, approximately 1 : 4.

It is interesting to notice that in the 37 years period occurred three times *maxima* below the 12 years annual average :

19 Feb. 1911		46 m ³ /sec.	(A — 15),
10 Feb. 1910		47 »	(A — 14),
13 Nov 1946		60 »	(A — 1).

Also the contrary happened, 1 *minimum* over 12 years annual average :

25 July 1915 85 m³/sec. (A + 24).

The diagram shows the monthly and annual average of the 12 years highest, lowest and (calculated) mean runoff. Our conclusion from the annual maxima and minima occurrence, formerly mentioned, that a normal water year culminates in January and reaches its deepest point in August is verified by the curves : but for the fact that in the 12 years period July may produce the minimum runoff rather than September. But the differences are very small ones only.

Coming back to our starting point, there can't be any doubt as to the motive of the scarce runoff of the last season. The 1946 maximum produced itself in November 13th with 60 m³/sec. only, 222 m³/sec. below the 37 years average (10). The snow melting concerns evidently the lowest parts of the mountains only. I have been told that Rio Blanco water didn't appear in the Rio de los Patos before the middle of December. The Rio Blanco is the most important tributary of the Rio de los Patos ; its affluent the Rio Colorado comes from the Mercedario group, the highest and most glaciated part of the whole share of our province in the Cordillera. Their water when at flood is very muddy, due to the glacial silt dragged along, hence the names of the rivers. Therefore Tamberías people know by the colour, when melting in this region has begun. Once more our rule concerning premature snow melting proved itself true.

San Juan, March 1947.

Published with consent of the authorities :
George J. Heinsheimer

(1) According to « El Régimen Pluviométrico de la Rep. Arg., published by the Dirección de Meteorología Geofísica e Hidrología, Buenos Aires, 1943, p. 33.

(2) Carlos Wauters, El Problema del Agua en la Región Árida de la Argentina, Buenos Aires, 1941, p. 142

(3) Statements in brackets added by the relator.

(4) A. de Quervain, Die Tätigkeit der Zürcher Gletscherkommission, Zt. f. Glk., 1921/2, p. 137.

(5) Carlos D. Storni, El Agua Subterránea en la Región Sudeste de San Juan, Buenos Aires, 1939, p. 99.

(6) Indeed according to a runoff diagram — see note 7 — drawn by the Dirección General de Hidráulica of San Juan, the increase of the river began in September and reached its maximum on December 28th., 1941 with 680 m³/sec, 298 m³/sec over or more than twice the 37 years annual maximum average.

(7) I wish to render here also my thanks to the Dirección General de Hidráulica de la Provincia de San Juan, for having liberally supplied the values of the annual maximum and minimum runoff 1910 to 1944 (different diagrams); monthly maximum and minimum runoff 1940 to 1946; monthly average stream flow 1935 to 1946; all these based on observations on the Dique Nivelador, Marquesado, Ciudad de San Juan. According to these data the tables 1 and 2 have been compiled. — The annual extreme values are referred to calendar, not water years; hence maxima and minima don't strictly alternate like in nature, but floods belong sometimes to the same runoff periods extending to the next calendar year. That is of course a source of errors.

(8) About errors resulting from referring the extremes to calendar years see note 7.

(9) Damian HUDSON, Recuerdos Históricos sobre la Provincia de Cuyo, Buenos Aires, 1898; Nicanor Larrain, El País de Cuyo, Buenos Aires, 1906; Augusto Landa, Resena Histórica de Obras Hidráulicas y del Régimen Legal de las Aguas en la Provincia de San Juan, San Juan, 1946; Augusto Landa, Cincuentenario de la Inundación de 1888, article in the San Juan diary « La Acción », January 1st., 1939. — I wish to render here also my thanks to the Ing. A. Landa, who not only indicated the literature mentioned, but furthermore kindly put his own personal annotations on the matter at my disposal.

(10) The dates concerning 1947 stream flow, I received after finishing this study are :

Month	highest	day	lowest	day	mean
47	daily runoff in m3/sec.				
January	51,306	2	36,782	21	41,288
February	44,410	5	34,730	28	39,725
March	34,894	5	26,096	29	30,249

Table N° 1

STREAM FLOW RIO SAN JUAN
Annual maxima and minima (Days average) 1910 to 1946

Year	Maxima		Minima		Year	Maxima		Minima	
	Day	m ³ -s.	Day	m ³ -s.		Day	m ³ -s.	Day	m ³ -s.
1910	10-2	47	27-10	14	1930	28-12	356	4-9	28
1	19-2	46	11-10	15	1	11-1	372	12-9	40
2	9-12	190	11-7	15	2	23-1	189	1-10	24
3	18-2	180	12-9	19	3	27-11	123	30-4	23
4	19-12	710	5-6	21	4	28-12	234	27-2	26
5	22-1	842	25-7	85	5	12-1	225	12-10	33
6	11-1	354	10-7	33	6	3-1	118	27-8	27
7	23-1	111	13-12	20	7	30-11	118	31-8	27
8	18-1	156	8-1	19	8	1-1	134	15-12	23
9	29-12	1096	17-7	30	9	14-12	93	27-8	22
1920	1-2	626	7-11	55	1940	27-12	151	7-4	23
1	17-11	306	1-5	50	1	26-12	680	12-8	28
2	1-7	232	4-10	44	2	1-1	504	10-8	53
3	28-11	120	1-11	28	3	6-1	196	22-9	40
4	1-1	107	6-12	14	4	29-11	216	12-8	35
5	27-12	172	5-9	22	5	17-2	140	12-8	26
6	7-12	509	22-5	26	6	13-11	60	20-9	28
7	1-1	350	15-9	39					
8	29-11	174	23-9	33					
9	29-11	145	16-8	27					

37 Years Average : Max. 282, Min. 30 m³-sec.

Table N° 3

RIO SAN JUAN — Distribution of max. and min. stream flow over the year 1910 to 1946

Months	Maximum stream flow				Minimum stream flow				Notes :	
	N ^o	Average	282 ±	Max.	Min.	N ^o	Average	30 ±		Max.
1	2	3	4	5	6	7	8	9	10	11
January	14	278	— 4	842	107	1	19	— 11	—	—
February	5	280	— 2	626	46	1	26	— 4	—	—
March	—	—	—	—	—	—	—	—	—	—
April	—	—	—	—	—	2	23	— 7	23	23
May	—	—	—	—	—	2	38	+ 8	50	26
June	—	—	—	—	—	1	21	— 9	—	—
July	—	—	—	—	—	4	41	+ 11	85	15
August	—	—	—	—	—	8	31	+ 1	53	22
September	—	—	—	—	—	8	31	+ 1	40	19
October	—	—	—	—	—	5	26	— 4	44	14
November	8	166	— 116	306	60	2	42	+ 12	55	28
December	10	419	+ 137	1,096	93	3	19	— 11	23	14

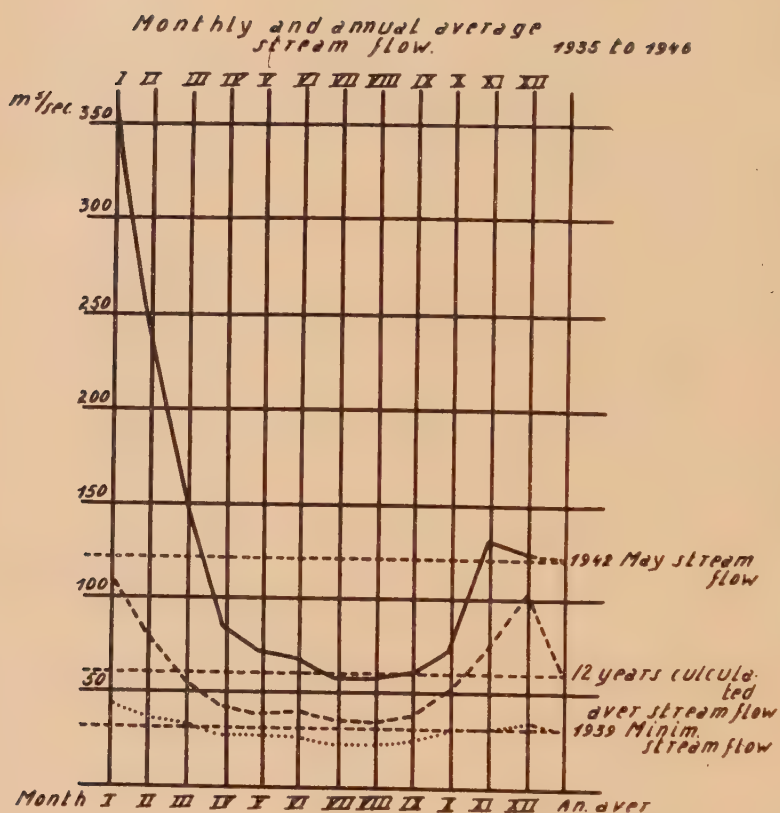
Column 3 to 6
and 8 to 11 : in m³-s.
Highest Max. : 1,096m³/s
Lowest » : 46 m³-s.
Ratio : 24 : 1.
Highest Min. : 85 »
Lowest Min. : 14 »
Ratio : 6 : 1.
Ratio Highest max. to
Lowest min. : 78 : 1
Average max. : 282 »
« min. : 30 »
Ratio : 9 : 1.

Table N° 4

Rio San Juan

**Relation between annual max. stream flow and the following annual min.
1910 to 1945**

Max. stream flow		Following minimum stream flow				R e s u m e
month	year	date		m ³ /s.	30±	
November	1921	4 X	22	44	+14	7 cases : 1 over average + 14 m ³ -s 6 below " — 6 "
	3	6 XII	24	14	—16	
	8	16 VII	29	27	— 3	
	9	4 IX	30	28	— 2	Months Febr., July, Aug. Sept. Oct. 1 each. Dec. 2.
	1933	27 II	34	26	— 4	
	7	15 XII	38	23	— 7	
	1944	12 VIII	45	26	— 4	
December	1912	12 IX	13	19	—11	10 cases : 6 over average +21 m ³ -s. 4 below " — 6
	4	25 VII	15	85	+55	
	9	7 XI	20	55	+25	
	1925	22 V	26	26	— 4	Months : April, May, July, Oct., Nov. 1 each August 2 September 3
	6	15 IX	27	39	+ 9	
	1930	12 IX	31	40	+10	
	4	12 X	35	33	+ 3	
	9	7 IV	40	23	— 7	
	1940	12 VIII	41	28	— 2	
	1	10 VIII	42	53	+23	
January	1915	25 VII	15	85	+55	14 cases : 8 over average +16 m ³ -sec 1 average " — 5 below average — 8 "
	6	10 VII	16	33	+ 3	
	7	13 XII	17	20	—10	
	8	17 VII	19	30	—	Months : August 2 July, Sept., Oct., Dic. 3 each.
	1922	4 X	22	44	+14	
	4	6 XII	24	14	—16	
	7	15 IX	27	39	+ 9	
	1931	12 IX	31	40	+10	
	2	1 X	32	24	— 6	
	5	12 X	35	33	+ 3	
	6	27 VII	36	27	— 3	
	8	15 XII	38	23	— 7	
	1942	10 VII	42	53	+23	
	3	22 IX	43	40	+10	
	February	1910	27 X	10	14	
1		11 X	11	15	—15	
3		12 IX	13	19	—11	Months : Aug., Sept., Nov. 1 each. Oct. 2.
1920		7 XI	20	55	+25	
1945		12 VIII	45	26	— 4	



11) OBSERVATIONS OF HIMALAYAN SNOWFALL AND SNOW-COVER MADE BY THE INDIA METEOROLOGICAL DEPARTMENT

I. Snow Observations

The importance of the snowfall and snowcover on the mountains to the north of India, especially during the winter months, for a proper understanding of the meteorology of India was recognised as soon as an All-India Meteorological Organisation was established in 1875, and since then attempts have been made to collect data regularly. Information is collected by getting monthly reports from District Officers, Forest Officers, Officers-in-charge of Mountain Passes, travellers, etc. Their reports usually contain information on (1) the number of days on which snow fall and a rough estimate of the quantity of snow fallen during a month, (2) the thickness of snow-cover on certain passes and stations at the end of a month, (3) the lowest altitude to which snow fall reached during the month and (4) qualitative comparison of the snowfall in the month with that in a «normal» year. This data supplied are inevitably in the nature of estimates and often incomplete, because the persons who supply the information as well as those who record them are not trained personnel. But however fragmentary may be the data for each year, as they have been collected from a wide area for a long series of years, it has been found that they yield useful information for various purposes.

2. In addition, the India Meteorological Department has been maintaining a number of meteorological observatories in the Hamalayas, where trained Observers record data of rain and snowfall along with other meteorological elements. A list of these observatories with their elevation and the year from which they are functioning is given below :—

Name of Station	Height above sea-level (ft.)	Year of starting
Drosh	4723	1897
Cherat	4272	1892
Murree	7113	1870
Gilgit	4890	1897
Srinagar	5205	1891
Skardu	7505	1896
Dras	10059	1896
Misgar	10190	1949
Leh	11529	1873
Simla	7224	1927
Mussoorie	6940	1929
Mukteswar	7592	1897
Katmandu	4388	1879
Darjeeling	7432	1867
Kalimpong	3965	1920

3. The information contained in the Snowfall Reports mentioned in paragraph 1 are scrutinised, consolidated and published in the annual « India Weather Review » every year. In this publication the whole range of the Himalayas is divided into seven regions and the available information is summarized separately for the four seasons as shown below : —

Regions

1. Afghanistan
2. Baluchistan
3. North-West Frontier Province
4. Kashmir
5. Punjab
6. United Provinces
7. Assam

Seasons

1. January and February (Cold weather period)
2. March to May (Hot weather period)
3. June to September (South-west monsoon period).
 - (a) June and July
 - (b) August-September.
4. October to December (Retreating monsoon period).

II. The Uses of Snow Observations in India.

4. Snowfall in the northwestern Himalayas during the winter season has a negative relationship with the subsequent monsoon rainfall (June to September) in northwest India. The physical basis appears to be that heavy accumulation of snow at the end of the winter months leads to a prolongation of winter conditions and delay in the establishment of the low pressure area over the North-west India, in proper form and intensity, so essential for the south-west monsoon mechanism. To work out the relationship, the information regarding snow accumulation in the northwest Himalayas at the end of May is converted into numbers by assigning marks for each year for each location for which the Report is available according to the following scheme formulated by Sir Gilbert Walker in one of his Memoirs on Correlation in Seasonal Variations of Weather. (1). —

<i>Character of snow accumulation</i>	<i>Marks</i>
Very small	— 2
Small	— 1
Normal	0
Moderate excess	+ 1
Large excess	+ 2
Very large excess	+ 3

The correlation coefficient between the snowfall accumulation at the end of May in the northwest Himalayas and the subsequent monsoon rainfall (June to September) in northwest India, computed for the period 1876 to 1940, is — 0.31. This is one of the most important factors, which are used for foreshadowing the monsoon rainfall in India.

5. A forecast of the precipitation for the region « Northwest India » (including the mountain plateaus and valleys) during the winter season has been issued by the India Meteorological Department for a number of years now. The significant factors which are used for the purpose are :—

Western rain (December) — Correlation coefficient + 0.36

South American Pressure (December) — Correlation coefficient + 0.41

Agra West upper winds $\frac{1}{2}$ (September II + October I) —

Correlation coefficient + 0.48

At present, this forecast gives the probable amount of winter precipitation (rain and snow combined, i.e., rainfall in the plains and snowfall over the mountain region) during the three months, January to March. Here also the data of snowfall on the hills is rendered quantitative by a system of markings as indicated below, also suggested by Sir Gilbert Walker (2).

<i>Character of snowfall</i>	<i>Marks assigned</i>
Phenomenally little snow	— 3
Much below normal	— 2
Below normal	— 1
Normal	0
Above normal	1
Much above normal	2
Abnormally heavy snowfall	3

6. The departures of snowfall for the period January to March and snowfall accumulation at the end of May, obtained by the above schemes of marking for the years 1876 to 1947 are given in Table I.

7. The effect of snowfall and snow accumulation in the Himalayas on the discharges of the Indus river system was studied recently with the available discharge and snow data. The correlation coefficient between the snow accumulation at the end of May and the Indus river heights at Bukkur, for the years 1876 to 1939, worked out to be $+ 0.37$.

8. Analysis has been made of the contributions by snow-melting to river discharges from available data. As an example Fig. I shows the Kosi Catchment and its sub-divisions and Tables II to V give some of the essential results of the computation.

III. Present Projects for improved Snow Observations and Snow-Surveys in the Himalayas

The need for more accurate and detailed information on the snowfall and snow cover in the Himalayan catchments to the north of India and for a systematic study of the factors in snow cover leading to runoff, has arisen recently in connection with the large scale water-power projects of the Government of India. The Central Waterways, Irrigation and Navigation Commission, which has been instituted by the Government of India to develop the water resources of the country, and the India meteorological Department are cooperating in setting up well-planned networks of hydrometeorological stations in the catchments of the more important Indian rivers. The plan for the Kosi river catchment, which lies in the eastern Himalayas between Khatmandu in Nepal and Darjeeling in India, embraces about 30 observation stations of which 20 will be equipped with raingauges and snow-gauges, 6 with snow-rain recorders, and 4 with other meteorological recording instruments. Of these, 9 snowgauge stations and 25 raingauge stations are already functioning.

Schemes for snow-surveying have also been receiving attention. March 1947, Dr. J. E. Church, President of the International Commission on Snow and Glaciers, came out to India at the invitation of the Indian Government to develop plans. He led parties of Indian Meteorologists and Hydrologists on short expeditions up the Teesta valley in the Sikkim Himalayas and up the Tamur valley in the Kosi catchment. A number of snow courses in the Teesta Catchment were laid out during these trial expeditions. In the winter of 1947, a party of Indian officers went again up the Tamur valley and laid a few snow courses in its upper reaches. In March 1948, a small party with two Nepalese Observers has set out to make snow surveys in these courses. Afterwards, the Nepalese Observers will remain at Wallanchangola to take rainfall, snowfall and river discharge observations at this place, and to make periodical measurements along the snow courses.

Meteorological Office,
Lodi Road, New Delhi (India),
the 18th May, 1949.

REFERENCES :

- (1) India Meteorological Department Memoirs, Vol. XXI, Part II.
- (2) India Meteorological Department Memoirs, Vol. XXIV, Part I.

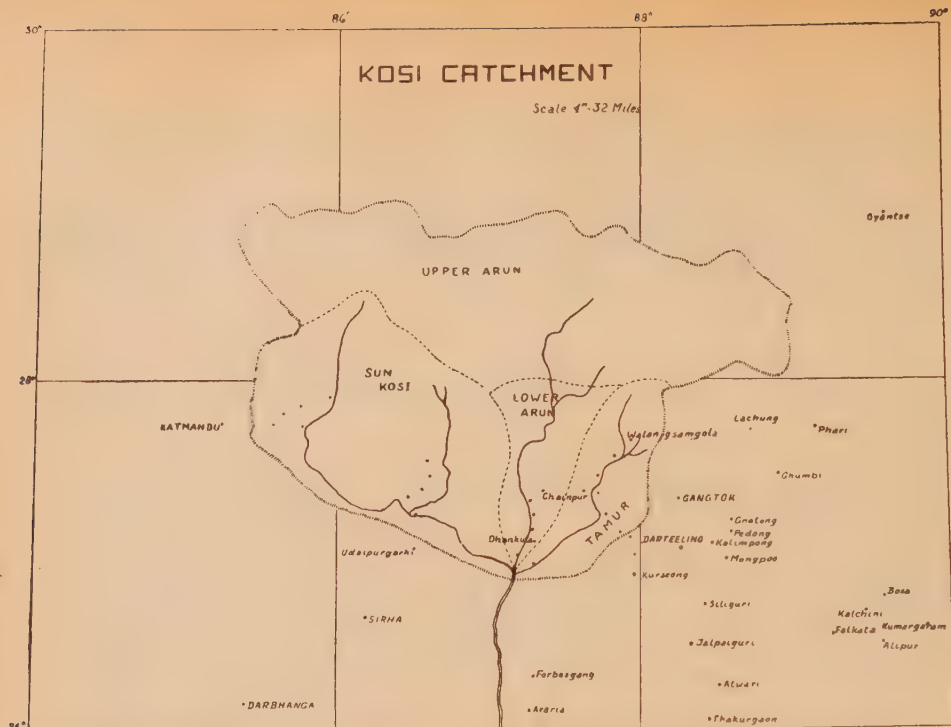


Table I

Departures of (1) Snowfall during January-March and (2) Snowfall accumulation at the end of May in the mountains to the north-west of India. (For explanation of numerical values see paras. 4 and 5 of preceding note).

Year	Snowfall Jany-March	Snowfall accumulation at end of May	Year	Snowfall Jany.-March	Snowfall accumulation at end of May
1876	— 1.0	+ 1.5	1912	— 2.0	+ 1.0
77	+ 3.0	+ 3.0	13	0	+ 0.5
78	+ 2.0	+ 3.0	14	+ 1.0	+ 0.5
79	— 1.0	— 2.0	15	— 1.0	0
80	— 1.0	— 1.0	16	— 2.0	— 1.5
81	— 1.0	— 1.5	17	— 2.0	— 1.0
82	— 1.0	— 2.0	18	— 0.5	— 0.5
83	0	+ 2.0	19	+ 0.5	+ 2.0
84	+ 1.0	+ 1.5	20	0	+ 3.0
85	+ 1.0	+ 3.0	21	— 2.0	— 2.5
86	+ 0.5	0	22	+ 0.5	+ 1.0
87	— 1.0	— 2.0	23	+ 0.5	+ 1.5
88	— 1.0	— 1.5	24	— 0.5	+ 2.0
89	— 1.0	— 0.5	25	— 0.5	— 0.5
90	— 2.0	— 1.5	26	1.5	+ 1.5
91	+ 3.0	+ 3.0	27	— 1.0	0
92	— 3.0	— 2.0	28	+ 1.0	0
93	+ 3.0	0	29	— 1.0	— 1.0
94	+ 1.0	— 1.0	30	0	+ 1.0
95	— 0.5	— 1.0	31	— 0.6	+ 0.5
96	— 0.5	— 2.0	32	— 1.3	— 0.5
97	+ 1.0	+ 1.0	33	+ 0.6	+ 1.0
98	— 2.0	— 1.0	34	— 1.0	0
99	— 1.0	— 1.0	35	+ 2.3	+ 1.0
1900	— 0.5	0	36	+ 1.3	— 0.5
01	+ 1.0	+ 1.0	37	0	0
02	— 2.0	— 1.0	38	+ 1.0	— 0.5
03	+ 0.5	+ 1.0	39	+ 1.0	+ 0.5
04	+ 1.0	+ 1.0	40	— 0.1	— 0.5
05	+ 3.0	+ 3.0	41	— 2.0	0
06	+ 2.0	+ 2.0	42	+ 0.8	+ 0.5
07	+ 0.3	+ 3.0	43	— 0.5	+ 0.5
08	— 2.0	0	44	— 0.5	0
09	0	— 1.0	45	+ 0.3	+ 1.0
10	0	0	46	— 1.7	— 0.5
11	+ 3.0	+ 2.5	47	0	0

Table II

Mean Precipitation over Kosi Catchment and its sub-catchments computed from Isohyets drawn with available normals.

Month.	Upper Arun	Lower Arun	Sun Kosi	Tamur	Kosi (whole catchment)
January	0.27	0.69	0.65	0.73	0.49
February	0.48	1.55	1.53	1.44	1.01
March	0.74	1.67	1.76	1.60	1.25
April	3.65	4.89	3.77	4.61	3.89
May	4.75	9.49	9.61	9.48	7.34
June	6.44	21.20	22.14	20.36	14.41
July	7.09	27.58	28.99	27.26	17.56
August	8.88	20.47	22.76	20.44	15.21
September	5.39	16.82	18.22	16.46	11.78
October	1.76	4.37	5.85	4.11	3.39
November	0.31	0.50	0.59	0.50	0.43
December	0.25	0.32	0.32	0.31	0.29
Total	40.01	109.55	116.19	107.30	77.05
Annual chart.	42.83	103.42	112.04	103.14	76.92

Table III

Normal precipitation in thousands of acrefeet for Kosi and its 4-sub-catchments together with precipitation of Upper Arun expressed as percentages of (1) Arun and (2) Kosi (whole catchment).

	Upper Arun (Snow melt)	Lower Arun	Sun Kosi	Tamur	Total (Col. 1+2+3+4)	% Col 1	% Col 1
	(1)	(2)	(3)	(4)	(5)	Col. (1+2)	Col. 5
January	164	78	271	96	609	67	27
February	292	176	638	188	1294	62	23
March	451	189	734	209	1583	70	28
April	2222	554	1571	603	4951	80	45
May	2892	1075	4005	1241	9213	73	31
June	3921	2401	9228	2665	18215	62	21
July	4316	3124	12083	3568	23091	58	19
August	5406	2319	9486	2675	19887	70	27
September	3281	1905	7594	2154	14935	63	22
October	1071	495	2438	538	4543	68	24
November	189	57	246	65	557	77	34
December	152	36	133	41	362	81	42
Annual	24358	12410	48428	14043	99239	66	25

Table IV

Showing the contribution of discharges from the Upper Arun catchment expressed as percentages of the discharges of (1) Arun and (2) Saptkosi. (Discharges in thousands of acre feet) 1947

	Arun discharges	Tamur discharges	Estimated discharges from Lower Arun alone Col. (2) × 0.865526*	Estimated contribution of Upper Arun to Arun discharges	Percentage contribution of Upper Arun to Arun discharges Col. 4 Col. 1 × 100	Total discharges Kosi In 1000 ^s of acre feet	Percentage contribution of Upper Arun to Kosi discharges Col. 4 Col. 6 × 100
	(1)	(2)	(3)	(4)	(5)	(6)	(7)
1947							
January	324	159	138	186	57	944	20
February	238	119	103	135	57	703	19
March	281	142	123	158	56	757	21
April	377	178	154	223	59	937	24
May	473	254	220	253	53	1232	21
June	1269	546	473	796	63	3364	24
July	2435	1708	1478	957	39	8828	11
August	2088	1869	1617	471	23	8702	5
September	1768	1289	1116	652	37	6655	10
October	1314	810	701	613	47	4235	14
November	492	355	307	185	37	1817	10
December	435	198	171	264	61	1263	21
Annual	11494	7627	6601	4893	43	39540	12

$$\frac{\text{* Area of Lower Arun catchment}}{\text{Area of Tamur catéhnent}} = \frac{2124}{2455} = 0.865526$$

Table V

Averages of monthly rainfall for Kosi catchment for 1947 together with monthly rainfall equivalents obtained from runoff figures and runoff ratios derived therefrom.

Month	Mean rainfall from isohyets 1947	Runoff in rainfall equivalents		Percentage runoff ratios	
	(1)	Saptkosi (2)	Total of sub- catchments (3)	Col. (2) Col. (1) (4)	Col. (3) Col. (1) (5)
January	0.10''	0.76''	0.74''	760	740
February	0.15	0.55	0.55	367	367
March	1.80	0.63	0.60	35	33
April	1.30	0.78	0.74	60	57
May	2.60	1.06	0.97	41	37
June	6.80	2.89	2.65	43	39
July	16.60	8.01	6.95	48	42
August	12.42	8.69	6.85	70	55
September	8.74	6.39	5.24	73	60
October	2.09	4.09	3.41	196	163
November	0.10	1.37	1.43	1370	1430
December	0.10	0.96	0.99	960	990
Annual	52.80	36.18	31.12	69	59

THE EVOLUTION OF SNOW-MELT BY DYES AND DRIP-PAN

by

J. E. CHURCH

For several years Fuchsine dyes have been in use at the Soda Springs, California (U.S.A.) Snow Station (1) by the author to trace the movement of snow-melt water. This was accompanied by measurements of temperature in the surface of the snow cover and at intervals above it. Then drip-pans with catch tanks were installed to determine the amount of the daily snow-melt and the time of its arrival at the base of the snow cover and the time of its diurnal cessation. Complete success attended the tank measurements at two stations (Fig. 1) tho the freezing of the valves at others (Fig. 2) prevented a satisfactory record.



FIG. 1.

Since February 1947, absence in the Himalaya and Andes has prevented the continuation of the studies, except the tank measurements, and no opportunity has been found to prepare the field studies for publication. Therefore, until the statistical details are available, the following conclusions are offered as a background for further study. It is really the study that should have preceded the first snow-survey forecast of streamflow, but as usual a practical thing becomes scientific only after our need and curiosity have impelled us to it. In this work, John Johansen has played the practical part, with Eleanor Johansen and myself assisting in the night measurements.

Our general observations and present conclusions are as follows. There will, of course, be minor variations and exceptions.

The fuchsine dye and snow sampler have made it possible to determine the finer points in the melting and freezing of snow. The thermometers and thermocouple have indicated the micro-climate of the snowfield.

(1) Conducted cooperatively by the Nevada Agricultural Experiment Station and the U. S. Weather Bureau

While the free water in the snow can be determined closely by weighing and by the dye possibly within 25 per cent of actual, the use of such measurements seems no longer necessary for the free water in the snow varies with the temperature conditions under which it is observed. The relative density of the snow or study of its crystal structure will indicate better its capacity to hold melt water. The feeling of Carol F. Merriam that there may be a temperature and density of snow at which melting can be explosive seems not in accord with the slow action of the dyes.

In the snow-melt season when the snow has become ripe or attained its highest capillarity, the snow cover protects itself from too rapid melting and evaporation by its own radiation. At midday the air temperature one foot above the surface has been found to be at 40° F., while at 5 feet overhead the shade temperature was 70° F. The air temperature immediately above the snow was thus restraining the rate of melting. Only when the warm wind blows over it or warm rain falls upon it will the rate be greatly increased and only when the sky is cloudy will the radiation from the snow cover be held to a minimum.



FIG. 2.

The mean melting of the snow during the month of April for several years at Soda Springs (see Meinzer's *Hydrology*, pag. 129) indicates that even at the mean temperature of 12.5° F. above freezing during the month, the melt water does not seem to have been above control by the soil. Perhaps this is the reason that temperature has so little effect on total runoff.

The snow therefore by its own radiation is its best protector. In the evening when the sun is getting low, if the weather is clear and the temperature at 5 to 10 feet above the snow is still at 40° F. or even higher, the surface crystals of the snow cover begin to freeze and the melting ceases. This has been noted not only at Soda Springs repeatedly but at timber line on Mount Washington and on Mount Rainer glacier by Max Demorest and on the Greenland Inland Ice. On Mount Washington

the temperature of 41° F. was scarcely five feet above the snow. In the Himalaya, April 7 (1947) at 12,550 feet where there had been rain and hail during the evening, the surface of the snow froze in the night during a brief period of clearing. This was shown by the change of the red dye to purple on the surface of the snow while it was red in the still wet snow underneath.

The percolation of melt water after surface melting has ceased continues only until the capillarity of the snow crystals resists the gravity load of the melt water. Then percolation practically ceases until the next day's surface melt begins, when the gravity load of the melt water quickly overcomes capillarity again.

The freezing at the snow surface is limited in depth and the snow beneath remains moist through dormant until the sun returns. The action of the dye upon the core of snow being studied together with pressure by the hand reveals the relative free water in the snow. It is interesting to plant the dye on the surface in the evening and at the same time at a depth of 6 inches or a foot beneath it to see whether the freezing of the crust can overtake the descent of the dye. With the head start of a few inches the dye will continue sometimes even to the bottom of the snow while the dye at the top penetrates scarcely one half inch. This indicates the continuing downward movement of the melt water until it has become so scanty that full capillarity has returned to the snow.

The depth of freezing at night does not seem to affect the rate of melting, for the next day's melt on the surface of the crust seems quickly to find its way thru fissures in the crust and make its progress as usual towards the bottom. In our test the red dye may have accelerated the surface melting somewhat. It is therefore best when possible to cover the dye with white snow in order to eliminate any artificial warming.

Of course, where the temperature remains high during the night the snow also continues melting. But in such areas there will be little holdover of snow for irrigation, for the elevation will doubtless be low. Furthermore, in a semi-arid region high rates of radiation and low night temperatures are prevalent in the mountains.

The melt day instead of being 24 hours long is more nearly 8 hours, with a period of dormancy in the snow approximately 16 hours. The melting does not start until the sun is well up and then only slowly, and it ceases while the sun is still above the horizon. The flow in the stream because of continuing percolation in the soil during the night does not cease entirely even at the snowfields.

The « degree days » should be based on the hours of melting, not on the mean daily temperature and the temperature should be measured at the snow surface and preferably by wet bulb thermometer. Only thus can field and laboratory measurements be compared.

To visualize the phenomenon of snow-melt it is planned to obtain a color movie of the snow crystals in the early morning as they burst into water under the rising sun.

13) **STANDARDIZING METHOD OF SNOW COVER
MEASUREMENT : TEN YEARS COMPARISON OF
FORECASTING RUNOFF OF THE COLORADO RIVER (U.S.A.)
BY SNOW SURVEYS AND MEASUREMENT OF
PRECIPITATION**

by

J. E. CHURCH

Summary of Forecasting Methods

Since the Colorado River on the U.S. Continental Divide is complex in its feeders and still depends for its forecasts upon precipitation and snow-surveys not yet organized by altitude zones, the analysis of snow cover and streamflow for the first 10 years of its snow-surveys is quoted (Tables 1 and 2) to indicate the details of the usual snow-surveys and their approximate accuracy.



The Colorado River that occupies the southern half of the western slope of the Continental Divide of North America is far overburdened with land to serve. Consequently, forecasting its water supply has become an earnest and delicate task. —

Fortunately, as compensation for its desert area, the Colorado's watershed of 225,000 square miles is reduced, for purposes of forecast by almost one-half and the crest line of 760 miles is diminished to 330 miles.—

Furthermore, three tributaries, the Green (5,797,760 acre-feet), the Upper Colorado (6,650,200 acre-feet), and the San Juan (2,745,270 acre-feet) furnish 87.1 per-

cent of the mean annual runoff of 17,489,000 acre-feet at Yuma. Moreover their runoff during April to July attains 67.0 to 75.0 percent of the annual, thus establishing their snow-fed character. The task of forecasting the runoff of the Colorado is one of concentration on the three main feeders mentioned.

In the case of a large and complex basin such as the Colorado, the harmonics of precipitation, snow cover, and runoff should be studied for the main stream and larger feeders. Data should be weighted by altitude zones as well as by relative contribution of subordinate areas to the total watershed. Search should be made for individual differences as a means of identifying and computing the factors bearing on runoff.—

The present analysis however covers only the precipitation and snow cover weighted by relative contribution of areas or feeders and the runoff of the Colorado at Bright Angel Creek near Grand Canyon, 267 miles upstream from Boulder Dam. The data for runoff have been expedited by the United States Bureau of Reclamation and the United States Geological Survey. The averages for snow cover and precipitation have been taken unmodified from the « Snow Survey and Irrigation Water Forecasts for the Colorado River Drainage Basin » by the Division of Irrigation, United States Soil Conservation Service, Table p. 146 (a), showing precipitation, snow cover, and runoff is based on the principles of forecasting gradually developed by the Nevada Agricultural Experiment Station. Some principles may bear repeating :

- 1) The major factor in the summer (April-July) runoff is the snow cover, which attains its maximum accumulation about April 1. Since the snow begins to melt steadily by April 1, the basic normal for computing percentage of the summer flow must be the snow cover on April 1 and not the snow residue on May 1. Any later readjustment in the forecast, as on May 1, must be a modification in the forecast of April 1 owing to departure from normal of the precipitation during the interval, as for example during April. The difference in the snow cover on May 1 from that on April 1 should be a combination of precipitation and temperature effect during April and may therefore differ considerably from the precipitation alone.—

- 2) The percentage of the precipitation during October to March of preferably November to March under ideal conditions of gage catch should correspond to the percentage of the snow cover providing winter melting has not occurred.—

- 3) The effect of soil moisture in the Autumn probably is small as in the Sierra Nevada watersheds because of the steepness of much of the slope, but should nonetheless be determined where possible. Departure from normal temperature during this period still appears to have little if any effect on total runoff, the obviously it does retard or accelerate it. There is finally the effect of diversions above points of stream gaging when the flow is considerably above or below normal. There may also be the effect of high or low watertable on the total flow, but because of the steep gradient this can in no wise resemble that of the Humboldt Basin, Nevada.—

- 4) Since the accuracy of early or long-period forecasts is being discussed, only one revision, May 1, is proposed, altho a second revision in June 1, would be closer to the actual since by this time approximately 40 percent of the flow during April to July has occurred.—

Table 1 presents data weighted but uncorrected, and the comparison is confined to April 1. Table 2, however, contains all available correction factors for both April 1 and May 1 and comparisons for both dates. The details precede the table.—

Details regarding data, Colorado Basin. — The means for precipitation and snow cover were obtained from the annual reports of Snow Surveys on the Colorado Basin, and the weights for relative contribution of water by the constituent areas were provided by the United States Bureau of Reclamation. No attempt has yet been made to segregate the precipitation and snow-survey stations by altitude in order to use altitude zones in addition to the relative areas of the principal tributaries. The normals for precipitation were derived from the United States Weather Bureau departures based on period of record. Those for snow cover and runoff are based fortunately on the same period, ten years, representing the extent of the snow survey. The temperature departure, provided by the United States Weather Bureau, represents the period of record.—

Accuracy of forecasting on April 1, Colorado Basin. — As already mentioned, the relatively heavy precipitation on the upper Colorado during April to July as compared with December to March (the period of accumulation) subjects the runoff period of April to July to the possibility of great diversions from the percentage

Table 1. Comparison of percentages of normal precipitation (October to March) and snow cover (April 1) with percentage of normal runoff, April to July, Colorado river basin (uncorrected)

I T E M	Normal	1936	1937	1938	1939	1940	1941	1942	1943	1944	1945
Precipitation, October-March, inch	5.8 ^a 6.46	73.0 ^a	105.2 ^a	118.5 ^a	104.0	90.9	133.8	132.2	110.9	92.0	96.7
Snow cover, April 1, inch water eq.	14.9	118.8 ^b	109.4	113.4	83.5	71.0	99.8	98.4	106.1	95.6	102.0
Snow cover, May 1, inch water eq.	12.0	126.3 ^b	89.2	105.9	100.4	105.3	148.4	104.1	55.7	121.8	131.1
Precipitation for April	1.1 ^a 1.3	40.1 ^a	31.2 ^a	96.5 ^a	53.6	124.2	219.8	172.9	51.4	133.7	90.3
Temp.-departure for April (Grand Junction), °F	52.4	2.8 ^c	-1.7 ^c	0.8 ^c	3.2 ^c	2.0 ^c	-3.2 ^c	1.8 ^c	7.9 ^c	-4.4 ^c	-4.6 ^c
Runoff of Colorado River, April-July, at Bright Angel Creek	9,006,000 ac-ft	96.0	97.8	125.0	63.9	51.4	133.7	127.8	85.6	114.4	90.2
Variation between precipitation and runoff	—	-23.0	7.4	-6.5	40.1	39.5	0.1	4.4	25.3	-22.4	6.5
Variation between snow cover and runoff	—	22.8	11.6	-11.6	19.6	19.6	-33.9	-29.4	20.5	-18.8	11.8

^a Represented only by Green River, Grand Junction, and Durango for the Green, Colorado above Palisade, and San Juan River Basins, respectively.

^b Incomplete.

^c Deviation from normal in °F.

indicated by the snow storage on April 1. The disproportion between the seasonal precipitation in the Sierra Nevada and Continental Divide is extreme. In the Sierra Nevada, the precipitation during December to March is 67.41 inches compared to an April to July normal of 15.17 inches, while on the Continental Divide the corresponding normals are 28.83 and 31.00 inches. It is, therefore, a source of high satisfaction to find that in eight years out of ten, the snow cover on April 1 with obvious correction that can be made at that time is within a maximum deviation of approximately 20 per cent of the runoff from April to July. This compares roughly with 10 percent in the Sierra Nevada. Remaining divergent years can be detected at the adjusted forecast of May 1.—

Available correction factors, Colorado Basin. — The correction factors available as of April 1 for modifying the percentage of precipitation or snow cover are (1) autumn priming before snowfall, (2) the height of the water table at the beginning of major runoff (April to July), (3) the disproportionate melting of the snow cover in the lower zone of the watershed in the winter, and (4) the effect of diversion above points of gaging. To these might be added the winter runoff as evidence of the effect of high winter temperatures.—

In the present analysis, corrections only for diversion effects have been made.—

Increase in accuracy of re-forecast of May 1: factor of precipitation during runoff, Colorado Basin. — In addition to the factors that can be recognized on April 1 there is also the factor of precipitation during runoff that can be recognized only as it occurs. Fortunately the normal precipitation for April is 29 percent of the total for April to July, and the precipitation for April to June, while the snow still covers considerable area, is 55 percent of the period of runoff. The monthly percentage based on joint weighted mean for Green River, Grand Junction, and Durango is 29.1 for April, 25.8 for May, 15.5 for June, and 29.4 for July.—

The precipitation for April to May is most effective and for April especially so, for practically the entire fall is caught by the snow cover which also provides the soil priming. Altho the precipitation in July is highest of all four months, its efficiency in producing runoff is probably least, owing to the large proportion of bare ground at the close of the melting season.—

Residual snow cover, Colorado Basin. — The snow survey of May 1 is planned to afford a view of the snow cover still remaining from the cover of April 1. It represents a combination of the melting that has occurred plus precipitation during the period. The net loss of 3 inches (14.9 inches on April minus 12.0 inches on May 1) (see Table 1) may represent a total loss by melting of five inches and an accretion of two inches of precipitation. Altho the weighted mean precipitation shown by the average of the group of precipitation stations, most of which are at lower altitudes, is 1.3 inches, the weighted mean of three representative high-level station is 1.93 inches.—

Correction factor for April precipitation, Colorado Basin. — On the basis of the high-level precipitation in April of 1.93 inches to the snow cover of 14.9 inches on April 1, the precipitation during April represents a factor of 13 percent in the runoff during April to July. The factor of correction for 100 percent departure in precipitation during April was set at 20 percent of normal runoff or higher owing to the large proportion of bare ground at the close of the melting season. But if the available percentages of 29.1, 25.8, 15.5 and 29.4 (at the low-level precipitation stations) for the four months are weighted by $\frac{4}{4}$, $\frac{3}{4}$, $\frac{2}{4}$, $\frac{1}{4}$ for decrease in snow cover area, they become 29.1, 19.4, 7.8, 7.4 respectively and April represents 45.7 percent of the total.—

Since the unweighted precipitation at 29.1 percent for April represents 13 percent of the runoff for April-July, the weighted precipitation at 45.7 percent should represent 28.4 percent. The factor of 20 percent had earlier been set on general considerations. The increase improves the factor proportionately.—

Corrected forecasts, Colorado Basin.— Table 2(a) indicates the corrections possible in the forecasts of April 1 and re-forecasts of May 1. The correction factors available are diversion April 1 and precipitation May 1. In a fully developed forecast, the other factors mentioned should if possible be included.—

Accuracy, Colorado Basin forecasts. — The accuracy of forecast on May 1 in the case of the snow cover is within 15 percent for the ten years of record and for four

years of record and for four years is within ten percent. If wholly uncorrected April 1, the maximum between snow cover and runoff for six years is within 20 percent and for the other four years is as great as 35 percent. For the precipitation, three forecasts for May 1 out of ten were within ten percent, but the other seven varied from 15 to 49 percent.—

Both the precipitation stations and the snow-survey courses should be arranged by altitude zones and more high-level precipitation stations provided. Yet the snow-surveys will provide a more reliable index of winter melting than departure from normal temperature recorded at a central station.

Parallel normals will assure more accurate comparisons. The normals of snow cover and runoff are both based on ten years' record. The divergence between these and precipitation may be due in part to this cause. The large divergence in 1941 and 1942 when the snow cover was approximately 35 percent of normal below the precipitation does not seem certainly to be due to winter melting if judged on the basis of temperature departure alone. Low-level snow-surveys and the measurements of the stream flow will be more conclusive.—

Further Refinement of Diversion Effects : Because of the diversity of snow cover and diversions on the individual feeders of the Colorado in 1946, a further effort was made to determine the probable residue or net runoff from each, whose combined runoff represents 90.5 percent of the main stream at Bright Angel Creek above Lake Mead.—

The snow cover this year for the Colorado when duly weighed by the tributary basins was 68.7 percent and the residue runoff, after 3,750,000 acre-feet (2.5 acre feet per acre) diversion, should be 55.7 percent of normal. The 3-acre feet per acre diversion was considered too high by George Malone because of considerable return flow and by H. P. Boardman because some of the diversion was made after April-July, the period which the forecast was made.—

The following table 2 contains the details of diversion effects by individual feeders :

Table 2. *Effect of diversions on net flow by individual feeders at 2-½ acre-feet per acre (Net Runoff of each on Basis of its Contribution to Total Runoff of Main Colorado above Lake Mead)*

Feeders	Percent	Snow Cover Acre-feet water	Diversions Acre-feet	Net Runoff Acre-feet	Percent
1. Green River					
Weight 26.5%					
Normal	100	4,267,280	1,592,500	2,674,780	100
1946	87.6	3,738,140	1,952,500	2,145,640	80.2
2. Upper Colorado					
Weight 40.0%					
Normal	100	5,375,590	1,485,000	3,890,590	100
1946	75.2	4,042,440	1,485,000	2,557,440	65.7
3. San Juan					
Weight 24.0%					
Normal	100	3,032,130	582,500	2,449,630	100
1946	36.0	1,091,570	582,500	509,070	20.8
Main Colorado					
Weight 100%					
Normal	100	12,756,000	3,750,000	9,006,000	100
1946	68.7	8,872,150	3,750,000	5,212,150	57.9
Total actual runoff main Colorado				5,561,300	61.8

The normal runoff of the main Colorado above Lake Mead is the 10-year mean of actual measurements near Bright Angel Creek, but the normals of the feeders, the Green, Upper Colorado, and San Juan, represent approximations based on the

Table 2(a). *Forecasts of runoff during April to July, Colorado basin above Grand Canyon, corrected for diversion April 1 and precipitation May 1, in percentage of normal.*

I T E M	1936	1937	1938	1939	1940	1941	1942	1943	1944	1945
Precipitation, October-March										
Corrected for diversions, April 1 ^a	73.0	105.2	118.5	104.0	90.9	133.8	132.2	110.9	92.0	96.7
Corrected for precipitation, May 1 ^b	59.5	—	127.8	—	86.8	150.7	148.3	116.4	—	—
Revised forecast, May 1	47.5	91.2	—	95.8	91.8	174.7	163.3	116.4	99.0	94.7
	47.5	91.2	127.8	95.8	91.8	174.7	163.3	116.4	99.0	94.7
Snow cover on April 1	118.8	109.4	113.4	83.5	71.0	99.8	98.4	106.1	95.6	102.0
Corrected for diversions, April 1	128.2	114.1	120.1	76.0	56.5	—	—	—	—	—
Corrected for precipitation, May 1	116.2	100.1	120.1	67.0	61.5	123.8	113.4	96.0	102.6	100.0
Revised forecast, May 1	116.2	100.1	120.1	67.0	61.5	123.8	113.4	96.0	102.6	100.0
Runoff at Bright Angel Creek, April-July	96.0	97.8	125.0	63.9	51.4	133.7	127.8	85.6	114.4	90.2
Divergence in revised forecasts, precipitation compared to snow cover	—68.7	—8.9	+7.7	+20.5	+19.9	+50.9	+49.9	+20.4	—3.6	—5.3
Precipitation compared to runoff	—48.5	—6.6	+2.8	+31.1	+40.4	+41.0	+35.5	+30.8	—15.4	+4.5
Snow cover compared to runoff	+20.2 (+14.2) ^c	+2.3	—4.9	+3.1	+10.1	—9.9	—14.4	+10.4	—11.8	+9.8

^a For method of computing diversion effects, see p. 263. Years within 10 percent of normal (the range of instrumental error) are not included.

^b See p. 264, departure 100 percent in April precipitation, departure 20 percent in April-May runoff. Example : In 1936, 60 percent of normal required a correction of 12 percent.

^c Incomplete snow survey for 1936 which affects both 20.2 and 14.2. With reference to 14.2, if correction is made for diversion on May 1 after the correction for precipitation departure has been determined, the snow cover of April 1 viz. 118.8 would be 106.8. The correction for diversion would then be based on only 106.8 and the corrected forecast would be 110.2 percent and variation from actual runoff 14.2 percent. This reverse method of waiting one month was suggested by the United States Engineers. It should give closer results. There would be no diversion effect in 1937 but the correction in 1938 would remain unchanged.

weights of 26.5, 40.0 and 24.0 percent, respectively, of the main Colorado increased individually by 3.2 percent to provide for 9.5 percent representing minor streams.—

The normal acre-feet water in the snow cover (column 2) is derived from the net runoff plus estimated diversion, but the percentage of the snow cover is based upon the snow surveys themselves.—

The computation of the net seasonal runoff is illustrated in the following forecast for the Green River 1946 : 87.6 % of normal 4,267,280 acre-feet — seasonal-3,738,140 acre-feet — diversion 1,592,500 acre-feet — net runoff 2,145,640 acre-feet or 80.2 % of 2,674,780 acre-feet normal net runoff.—

The estimated net runoff based on the net runoff of the individual feeders was 57.9 percent in place of 55.7 percent estimated for the basin as a whole or a gain of 2.2 percent.—

The divergence between the former or refined estimate and the actual runoff of 61.8 percent is only — 3.9 percent : too close for acceptance as fact.

The Weak Efficiency of Summer Precipitation : The uniformity of the major runoff phase of the western streams of North America in terms of the annual and its restriction to April-July, or with latitude or altitude to May-August, indicates that snow is the dominant factor irrespective of summer precipitation.

This is shown in the following comparison (Table 3) of the runoff phases of the San Joaquin, Upper Colorado, and Bow Rivers, which represent a cross section of the mountain west. The first is in the Sierra Nevada and the others are on the two sides of the Continental Divide.—

Of the three streams there is a tendency for the major runoff of the San Joaquin to begin in March whereas the Upper Colorado begins only in April and runs over slightly into August. The northern Bow begins only in May and continues into September.—

Irrespective of the disparity in total precipitation in the basins, the percentage of the major runoff in terms of the annual is high being 70.10, 75.04, and 68.4 for the three streams.—

The efficiency of rainfall as compared with snowfall depends primarily upon the presence of a snow cover upon which to fall and thereby escape the priming loss that occurs when it falls on soil. For example, in the Susquehanna Basin of the Appalachians, although snow constitutes one-third of the annual precipitation, it is far more effective than the remaining two-thirds. In like manner the 28.83 percent of December-March precipitation at Grand Junction in the Upper Colorado-Basin will be far more effective than the other 61.78 percent, but will naturally be aided by the 31.0 percent during April-July so far as the latter falls on snow.—

The April precipitation in the Colorado appears therefore to have a maximum effectiveness of 20 percent on the total April-July flow, but the May-July precipitation only 15 percent. Thus on May 1 the revised forecast based on the April 1 snow cover and April precipitation should be accurate within 15 percent of normal, as has been found true in each of the past 11 seasons. This has also been found true even on April 1 in 6 seasons or a majority of the total.—

However, the streams on the eastern slope of the Continental Divide may require a larger correction factor for April (or May) precipitation, but the character of the vegetation and sparseness of the total April-July rainfall (Banff May-August 11.19 inches) indicate that the precipitation will add little to the flow after the snow has largely disappeared.—

The evidence of the vegetation seems convincing. As one travels westward over the Great Plains, the grass uplands of eastern Wyoming give way at the crest of the Big Horn Range to sagebrush that prevails into the high foothills of the Continental Divide, in like manner as sagebrush grows high up the eastern slopes of the Sierra Nevada where the summer rain is also light. A comparison of the normal April-July precipitation at Caspar and Sheridan (7.06 inches and 7.82 inches respectively) east of the Big Horns and at Cody (4.74 inches) on the west indicates a sharp decline in precipitation toward the Continental Divide. This decline holds good also for the normal annual precipitation, viz. 14.58 inches, 15.06 inches and 8.98 inches. —

This picture increases confidence in the stable forecasting of streams to the west of the Continental Divide and affords hope that the streams immediately to the east will be found to flow in fair harmony with the snow at their source.—

Table 3. *Uniformity of major runoff phases in mountains of the west irrespective of summer Rainfall :*

	SAN JOAQUIN	UPPER COLORADO	BOW
	Runoff	Runoff	Runoff
	Percentage of	Percentage of	Percentage of
	Annual Friant	Annual Fruita	Annual Calgary
	Ann. 2,053,500 A.F.	Moab, Cisco	Ann. 2,704,150AF
	April-July	Ann. 7,540,000AF	May-August
	1,439,800 A.F.	April-July	1,850,000 A. F.
		4,850,000 A.F.	Approximate
October	1.48	4.05	5.9
November	1.35	2.64	3.9
December	2.30	2.02	2.7
January	5.96	2.04	2.6
February	5.00	2.06	2.2
March	8.07	3.37	2.4
April	13.48	8.12	3.3
May	20.37	21.22	8.4
June	24.62	32.07	22.3
July	11.63	13.63	23.0
August	3.69	5.77	14.6
September	2.06	3.02	8.6
April-July	70.10	75.04	May-August 68.4
December-March	21.33	9.49	Dec.-April 13.2

Improvements in Snow Cutters and Seasonal Gages

The Walf Tooth snow cutters (Fig. 1) (c) and (d) now makes it easily possible to cut the heavier ice strata in a snow cover.

The depth of ice that can be penetrated, however, is limited by the lack of storage space for the ice shavings can not be compacted into the ice being penetrated as is true in the case of snow.

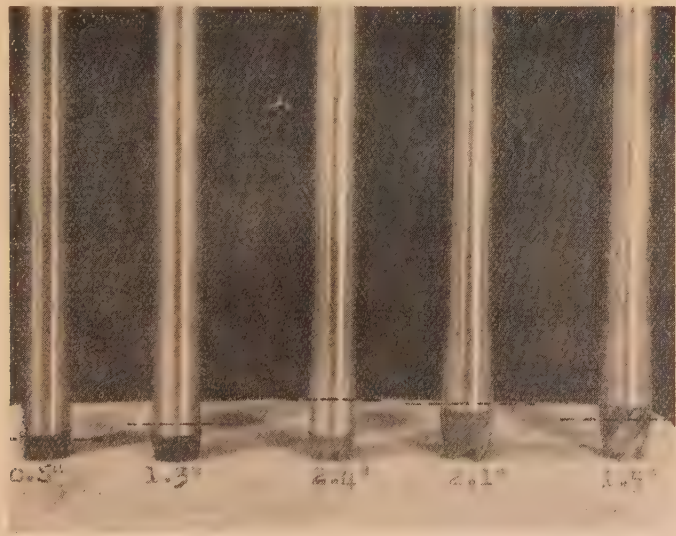


FIG. 1.

Ice penetration with one and the same man driving five minutes.
a) Now provided in current samplers.

- b) Original Mount Rose Cutter.
- c) Walf-tooth Cutter — Every other tooth cut out.
- d) Improved Walf-tooth cutter but screwed into an inverted coupling for easy removal.
- e) Ice auger. Teeth evidently so wide that the ice shavings are in excess of space to store them, The same may be true of (d). Cutters (c) (d) (e) were designed by John T. Ryan of the University of Nevada (U.S.A.)

Gages

The Sacramento seasonal Snow and Rain Gage (Fig. 2) designed by the U.S. Engineers and U.S. Weather Bureau has the advantage of large storage and a flat bottom to facilitate direct stick measurements.

The open wind-shield affords opportunity for the passage of dry snow and yet will close up on the windward side in heavy wind and block the upward current of air.



FIG. 2.

Occasionally in heavy snow precipitation (Fig. 3) both the orifice of the gage and the shield may become loaded with adhering snow which will persist until the sun returns and smelts it loose.

In the illustration the cloudiness continued for a week and caused a loss in the record of 10 percent of the normal annual precipitation.

Only the presence of an observer to clear away the snow can prevent such loss. Since the snow is frozen to the metal, wind action will rarely release it, but the use of dull black paint will intensify the action of the sun and accelerate final clearance. Since an airtight surface over the contents will prevent practically all evaporation either winter or summer precipitation gages need no longer be painted white to reduce heating.

It will be noticed that the snow accumulation on the alter or open shield is entirely separate from the snow cap on the top of the can, In the case of the light

shield of the Alpine type, any snow accumulation can gradually increase until it covers the top of the gage. However, the efficiency of the alter shield could probably be increased by placing the vertical slats closer together as in Fig. 2.

The weakness of precipitation gages is in the fewness of their number as compared with snow-sampling and the difficulty of catching snow.

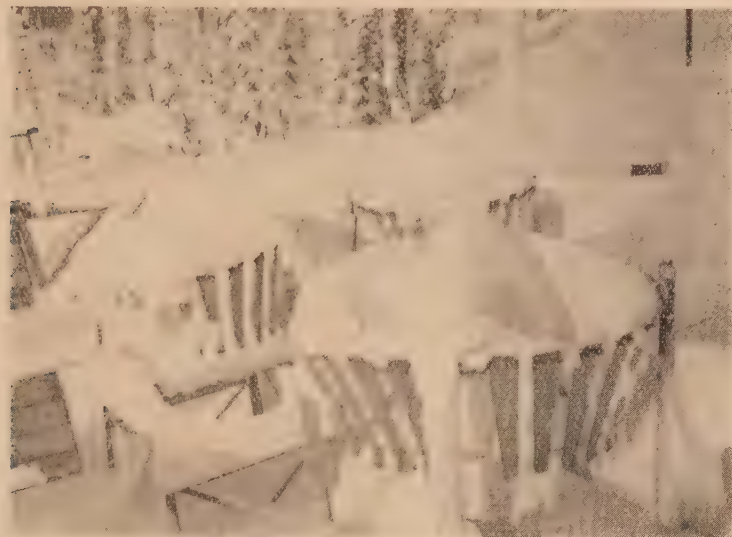


FIG. 3.

But more important still is the fact that gages can indicate only the total precipitation of the winter whereas snow-sampling indicates the snow cover residue at the beginning of the snow-melt season upon which the flow and the forecast must depend. The seasonal gages by capping may fail even more.

In the comparison in the Colorado River Basin the forecast based on approximately twenty gages, most of them read daily, attained the accuracy of the forecast by snow-surveying only six times in eleven.

However, precipitation gages are essential at the snow fields to show the precipitation that falls on the snow cover during the period of melting and for convenience should preferably be seasonal gages.

14) CANADIAN INTEREST IN SNOW AND ICE RESEARCH

by

R. F. LEGGET, Director, Division of Building Research

The Dominion of Canada stretches between two oceans, the Atlantic and the Pacific, from the borders of the United States of America as far south as latitude 41° to the North Pole. It extends over 48° of latitude and 84° of longitude. Within these bounds, it has an area of about 3,700,000 square miles, thus being the largest country in the western hemisphere and the third largest country in the world. The whole of this vast region is covered with snow for at least a short time during every winter; well over ninety per cent of the area is blanketed with snow for a winter period of several months. Despite the fact that over a third of Canada is still covered with forests, and that more than one half of its area is accounted for by the barren land of the Canadian North, the importance of snow and ice to Canadian living can readily be appreciated. That importance is at last being given recognition; this paper is an attempt to summarise this awakened interest of Canadians in the materials which are the concern of the International Commission on Snow and Glaciers.

The development of Canada as a modern agricultural and industrial nation has naturally been conditioned by its rigorous winters. In most parts of the country, agricultural work is possible for little more than half the year. This gives rise to an unusual intensity of activity during the short growing season and is one of the major causes of that «seasonal employment» which is so marked a feature of the Canadian economy. Until relatively recently, long distance travel during the winter by any means other than by railway was almost unknown, save only in the far reaches of the North. Here winter travel is essential, being often easier than at any other time of the year. Records of the endeavours of the early explorers of the North and the unimaginable hardships to which they were exposed have paved the way for widespread appreciation of the exploits of such men as those on the staffs of the Hudson's Bay Company and the Royal Canadian Mounted Police in their annual winter travels.

It is only about ninety years ago since these words were penned by a serious student of Canadian affairs: «We can scarcely credit the story that a railway between Quebec and Halifax is seriously contemplated. We are satisfied that no sane man either at home or abroad would invest one brass farthing in such a concern... Were a snowstorm to set in and it be as it is on our well-conducted roads, impossible to go forward or return, the passengers in the cars would most certainly be devoured by wolves or frozen to death... Great undertakings may be very captivating to men of enlightened minds and visionary ideas, but it is only fools who would embark upon them.» (1). «Fools» did embark upon them, to their credit and to the benefit of Canadians of today. They tackled the problems of the snow, the difficulties of ice, and the varied complexities of winter conditions with the result that road rail and air services today operate throughout the Dominion the year round. The water power plants which serve all parts of the country never stop; water supplies are perennial.

In meeting the problems of snow and ice, Canadians have made an approach which has been almost exclusively empirical. A great volume of sound knowledge has been accumulated by men who did not fear to experiment. Tribute to this practical approach must be a leading statement of this paper for it is upon the foundation thus provided that scientific enquiry can alone advance. A start has been made at this application of science. Snow and ice are at last coming to be recognised as materials as complicated as they are widespread, well worthy of detailed and meticulous study.

Hydrological Aspects of Snow and Ice :

Since all but a negligible proportion of the centrally generated electricity used in Canada is derived from water power plants, it is not surprising that early attention was paid to the possibility of forecasting stream flow from winter studies of snow and ice. The natural resources of the Dominion are administered by the several provinces, control having passed to them at different times from the central govern-

ment. Both federal and provincial organisations are therefore concerned with these snow and ice studies. The Dominion Water and Power Bureau works in close association with the provinces, its activities being nation-wide; its estimates of run-off are made in conjunction with the U.S. Geological Survey. It operates snow survey courses on the eastern slopes of the Rockies, obtaining accuracies of about 90% to 110% in this western area. In eastern Canada, similar work is done but in the Lake of the Woods area, for example, errors sometimes of the magnitude of 40% are found in run-off predictions for reasons not yet explained.

As an example of a provincial service there may be mentioned the snow survey work of the Division of Water Rights of the Department of Lands and Forests of British Columbia. This system has been developed since 1935 and now consists of 48 courses extending across the southern part of the province, from the coast to the summit of the Continental Range and as far north as latitude 53°. The courses are tied in with the extensive network of similar snow surveys in the western United States, the work of which is so well and widely known. In some areas, forecasts of spring run-off are consistently good with annual errors varying from 1% to 10%. In others, similar satisfaction has not yet been obtained. Study proceeds as to the causes or error, possibilities under consideration including :

a) The « soil priming » factor in watersheds the surface of which consists largely of soil;

b) Snow evaporation during the melting period, particularly in areas subject to the famous « Chinook » winds;

c) The effect of forest cover.

Similar snow survey work is carried out in Ontario by the Hydro Electric Power Commission of Ontario, in conjunction with the Dominion Water and Power Bureau. In Quebec the Shawinigan Water and Power Company has conducted similar investigations since 1928 and the Quebec Streams Commission is now extending this work. It is certain that the importance of this type of snow investigation, now so well recognised generally, will inevitably lead to its extension and further correlation across the Dominion.

It is somewhat surprising that the same attention does not yet appear to have been devoted to the hydrological aspects of snow in relation to agriculture. The Dominion Meteorological Service and other agencies regularly measure and report snow depths at many points across the country, but the detailed study of snow depth in relation to ground infiltration has yet to be fully developed. Possibilities of snow as a fertiliser have been suggested in a paper by Shutt to the Royal Society of Canada (2) but no extensive investigation has yet been initiated. Experiments have been carried out, however, on the control of snow drifts in order to increase the depths of snow on fields in order to augment ground water. This investigation and similar work has been undertaken by the federal Department of Agriculture.

More recently, renewed attention has been given to ice, in the form of glaciers, as a source of stream flow. The popular belief that the glaciers on the eastern slopes of the Rockies are « retreating » is now under scientific observation. The paper presented to this Conference by Mr. V. Meek, Controller of the Dominion Water and Power Bureau, summarises the present status of this important work. (3).

These several aspects of the hydrological value of snow and ice studies are well recognised; the surveys and practices described are not peculiar to Canada. They have served usefully to direct the attention of scientists towards winter conditions and, at least to some degree, are linked to broader and more general interests in snow and ice now to be described.

Ice Problems :

The rivers and lakes of Canada provided almost the only transportation routes for the first two hundred years of the country's development. They are still important arteries of transport, the Great Lakes and associated waters constituting the greatest system of inland waterways in the world. It has been asked « If the invention of the thermometer had preceded the discovery of the St. Lawrence, would the white man have gone back to Europe ? » — a hypothetical and scarcely accurate bit of rhetoric which epitomises the rigorous effect of winter on this transportation system. From the very earliest days of settlement, the coming of the ice and its « break-up » have provided annually two serious interruptions to the use of rivers and lakes. To meet these difficulties the famous amphibious « ice-boats » of the St. Lawrence were developed and are still in use (4).

Once the ice has formed, rivers and lakes again become traffic arteries, although now for surface vehicles rather than boats. Four inches of ice have come to be regarded as the minimum thickness necessary for the safe passage of a horse with its loaded sleigh: thereafter, until the coming of spring, the loads transported over ice are usually limited by the size of vehicle rather than by the strength of the ice. In the spring of 1948, for example, valuable mechanical and electrical equipment, weighing in all 220 tons, was safely transported by tractor trains for a distance of 325 miles over the ice of Great Slave Lake on its way to a far northern water power site.

With increasing use of ice for such transportation purposes, including the now widespread use of ice for the landing of aircraft, empirical methods for the determination of ice thickness and strength are ceasing to be adequate. Study is therefore being given to this complex problem, or group of problems. Many devices have been developed for rapid determination of ice thickness notably by the National Research Council, using such modern methods as supersonic and electronic techniques, but so far without real success. The problem is one of unusual difficulty but of comparable importance.

A theoretical analysis of the strength of a floating ice sheet has been completed but not yet published. Much experimental work can even now be envisaged but little has been done except in connection with the « Habbakuk » project of the recent war. Although no Canadian reports on this have yet been released, so that no reference to them may be made in this paper, a general account of the schema has recently been published by Perutz (5). A floating aerodrome was visualized, for Atlantic use, constructed of ice. An extensive programme was carried out, much of it at Jasper Lake in the Canadian Rockies. Laboratory work had been started by the British Department of Scientific Research (Admiralty) and was later continued in Canada under the auspices of the National Research Council to the direction of Professor G. M. Williams. The development of « pycrete » (a combination of ice and wood pulp) is described by Perutz in his paper: this interesting material is suggestive of other possible avenues of research.

It was not inappropriate that the « Habbakuk » work should have been done in Canada in view of the pioneer work on the properties of ice carried on for many years by Dr. H. T. Barnes of McGill University. In his presidential address to Section III of the Royal Society of Canada in 1909 (6), and later, in his well known textbook « Ice Engineering » (7), Dr. Barnes brought together a remarkable collection of information, supplemented by the results of his own extensive experiments, some of which are still remembered around McGill in almost traditional manner in view of their unusual character.

A prime interest of Dr. Barnes was the prevention or the alleviation of the « ice-jams » which occur in most of the rivers of Canada when the ice begins to move in the spring. Dr. Barnes did much work, and achieved some success, with the use of thermite for breaking up the ice packs, particularly in the St. Lawrence at Montreal. Flood conditions due to ice packing are sometimes so serious in that vicinity that the harbour is still surrounded by a massive protection wall, the gaps in which are annually closed at times of flood danger. Much more serious are the annual flood conditions at the scattered settlements along the Mackenzie River system (Canada's largest river), due to its flow from south to north. The thawing of tributary streams before the main stream has broken up gives rise annually to floods of such proportions that in some areas, no buildings or equipment can safely be left within sixty feet above the normal water level; they have to be moved up the steep banks in the fall and replaced after the flood has subsided.

Despite the work of Dr. Barnes, the problems of dealing satisfactorily with ice jams on Canadian rivers remains. Possibly the most dramatic evidence of the force of ice, when thus constrained, was the destruction of the famous « Honeymoon » bridge at Niagara Falls in January 1938 (8). This steel arch bridge, with a span of 1244 feet, was built in 1898 with its supports presumably fixed well above any possible flood water or ice level. Despite its successful performance for forty years, the combination of wind, temperature and ice conditions in the spring of 1938 on the Niagara River caused such a blockage of ice in the constricted gorge below the Falls that the level of the ice at the bridge site gradually rose until it was above the level of the arch supports. As the ice moved slowly downstream, the great bridge was pushed off its bearings as if it were matchwood, collapsing on to the ice and eventually disappearing into the depths of the river. The corresponding rise in water level did much damage to adjacent power houses, one being completely

flooded and filled with large ice blocks, necessitating months inactivity while repairs were effected.

This was an unusual mishap for water power plant but until comparatively recent years other power plants (and water system intakes) have been much troubled with « frazil ice » — the thin needle — like form of ice which will form in large quantities under certain conditions of super-cooling. Much investigation of this problem has been carried out in Canada, notably by Mr. John Murphy. The result of this work has been the development of design features for water intakes of both power and water supply plants which satisfactorily obviate the blockages so often in the past by « frazil » (9).

Another problem which arises in connection with Canada's water power plants is the determination of the amount of ice pressure to be allowed for in the design of dams and other water retaining structures. When it is noted that the figure, generally used is 10,000 pounds per foot of dam, exerted at maximum water level the economic significance of this problem will be at once apparent. Again, the problem is complex with the result that little study has been made of it. One of the very few fundamental approaches to the problem is provided by the work of Brown and Clark who carried out some singularly interesting experiments in a Montreal cold-storage warehouse in connection with the dam design for one of the Dominion's most northerly power plants, that at Island Falls on the Churchill River in northern Saskatchewan (10). It would appear that no accurate field measurements of the pressures actually exerted by ice have yet been made. A special committee of the American Society of Civil Engineers has recently been reconstituted to deal with this problem. It is hoped to conduct experimental work in Canada, in the field and in the laboratory, in the near future in conjunction with the current construction programme of the Hydro Electric Power Commission of Ontario.

Ice pressure is of similar importance in connection with the construction of marine structures on the east coasts of Canada, many of which are ice-bound for a portion of each winter. This freezing of the sea approaches to the Dominion has led to the development of a powerful Canadian fleet of ice-breaking vessels, now under the federal Department of Transport. First in the St. Lawrence and later in the Hudson Bay area, these sturdy vessels assist the spring « break-up » with a view to speeding up the opening of the navigation season. Design of the most recent vessel for Canadian winter service — the « Abigweit », the largest ice-breaking train-ferry vessel ever to be built — brought to light the dearth of information regarding the design of vessels intended for ice breaking (11). No record appears to exist of any approach to the problem other than that of experience; in view of the forces involved, and of the economics of the problem (the « Abigweit » for example has a displacement of 7600 tons), it would appear that here, also, is a field for research yet unexplored (12).

No review of Canadian ice problems would be complete without a brief reference to a singularly informative report by Kerry (13). A case is made out in this paper for further study of the temperature variation in the waters of the Great Lakes and the St. Lawrence in view of possible improvements to the navigation season suggested by the fact that water has its maximum density at 38° F. The paper, and the relevant discussion, provide a most useful fund of information and are fruitfully suggestive of yet another direction in which Canadian ice investigations may be expected to advance.

Snow Problems :

The problems to which snow gives rise are mainly related to the fact that some of it falls where it is not wanted and has to be removed. Its deposition can not yet be controlled. The invaluable records now provided by the Dominion Meteorological Service make it possible to predict local snowfalls within a fair range of accuracy. There is always the possibility, however, of a fall such as that which occurred in the Toronto district on the night of December 11th 1944 when 33 inches fell during less than twenty-four hours. The entire metropolitan area of Toronto, with its million people, was immobilised for a full day and all normal city services were disrupted for more than a week. Fortunately such occurrences are rare but when they do happen, they demonstrate in no uncertain manner the dependence of modern means of transportation upon snow clearing equipment.

Even the smallest Canadian municipalities now maintain some snow clearing equipment, often operated in connection with other automotive equipment of more

general use; the larger cities have large investments in their special snow clearing machinery. In all but exceptional cases, such as that just described, snow falls are cleared almost as rapidly as they occur. Indeed, in the city of Ottawa it was charged last winter that snow was cleared too quickly and too efficiently with the consequence (it was alleged) that frost was allowed to penetrate beneath city streets to a depth which endangered water supply pipes. This matter is to be studied in detail. Disposal of the snow is often a problem. In some cities it is dumped on vacant land; in others, onto local lakes or frozen river surfaces. In Montreal, manholes are specially located so that snow may be dumped directly into them, the heat of the sewage and other waters carried by the sewers being sufficient to melt the snow and carry it away as water to the out falls in the St. Lawrence.

Correspondingly, all major highways, and many strategic minor highways, are kept open the year round by the various provincial highway departments responsible (the federal government being charged only with highway work in the North-west Territories and the Yukon). The efficiency of this highway clearing is remarkable. In recent years, the province of Ontario has been materially assisted in its snow clearing work by a special weather predicting service provided by the Dominion Meteorological Service, which is passed on to highway users. The costs of winter road maintenance however, are very high. Taking the province of Ontario as an example, three million dollars is the average current expenditure for one winter. This covers only the cost of snow removal and the «skid-proofing» of roads by the use of sand (200,000 cubic yards a year) and special chemicals (40,000 tons). Already the supply of suitable sand is critical and the damage done to vegetation, to concrete pavements and to the undercarriages of automobiles by chemicals is serious. Here are two «minor» problems demanding attention.

It is difficult to prepare any accurate estimates of the cost of snow removal in Canada especially in view of the variation of climatic conditions with locality and time. Ontario's expenditure is typical. Montreal and Toronto each spend about one million dollars every year. The total cost, to provinces and municipalities, certainly exceeds ten million dollars each year and may well be considerably higher. A comparable figure is the total of twenty-one million dollars for the thirty-six States (of the U.S.A.) in what is called the snow-belt; it is to be noted that their climatic conditions are not nearly so rigorous as those of the Canadian provinces.

These vast sums are well and carefully spent. It can not be suggested, however, that the efficiency of snow moving equipment is as high as it might be. The possibility of some improvement in the handling of this one snow problem alone would seem to justify an extensive research programme. There are, however, comparable problems of almost equal magnitude. For example, the figures quoted do not include the cost of providing and installing the many thousands of miles of «snow-fencing» which are so familiar a feature of the Canadian winter landscape, by the sides of roads and of railways. The location of these fences is again a matter which apperas to be decided entirely empirically. Many fences are probably well placed but the researches of Finnie, in Michigan, have indicated the great possibilities for improvement (14). The problem is one involving aerodynamics; the first steps in treating it accordingly have been taken in Canada and, in conjunction with field tests, they indicate already some interesting possibilities.

The railways have been mentioned; they have their own snow clearing difficulties. Although their costs for winter operations are not published, they are known to run also into millions of dollars. In level areas, the problems are similar to those on highways; so also is the equipment, although somewhat heavier. In mountainous regions, special problems are encountered; rotary plows have to be used, and in some locations special snow sheds erected to obviate dangers from avalanches. In yard areas, and at busy terminals, critical problems, may often arise in the maintenance of switches during bad storms. In the Montrael terminal area, for example, over one thousand men may be employed during a storm merely for maintaining switches. And the special problems caused by snow drifting in cuttings and depressed areas are well known if only because of the publicity which they attract, as when a train in Saskatchewan was «snowed in» completely for over a week during the bad winter of 1946-47.

Air transportation too, has its special winter problems. There is the very technical matter of the icing of propellers and of wing tips upon which much research has been carried out in Canada, with attendant success. Airports have also to be maintained, bringing their own special snow-clearing problems. Large snow-blowers are now regularly used for this purpose, the practice being merely an extension of highway clearing. Alternatively, the practice of rolling and compacting the snow

has often been adopted, especially for smaller and more isolated airposts. The idea is a simple one; its effective prosecution carries many difficulties with it, some of which are still unresolved. When it is noted that Canada maintains over one hundred civil airports, the magnitude of this problem and its economic significance will be apparent.

Another associated problem is the use of skis in place of the usual wheels for the landing of aeroplanes on snow and ice. For light planes, skis are practically standard winter equipment. Their proper design became a matter of concern to the National Research Council in 1934 and Mr. G. J. Klein was assigned the task of investigating the problem. Many discouragements were met with in this work in view of its complexity but Mr. Klein was able to present a review of his findings at the 1939 meeting of this Commission. His work has now been described in a published report summarising an acceptable theory which meets the many facts which have been recorded (15). The work still goes on, its continuation over more than a decade indicating the complex nature of the problems involved.

The problems presented by aircraft skis are not dissimilar from those involved in the design and operation of sleighs for over snow transport, possibly the most wide-spread and amongst the most important of all Canadian winter problems. While it is true that major highways and all railways are now kept open throughout the winter, probably ninety per cent or more of the total mileage of Canadian highways are not so maintained for winter travel, thus necessitating the use of over-snow vehicles. Sleighs still predominate in this field. In logging operations, they are of cardinal importance so that it is not surprising that one of the first Canadian studies of the over-snow transport problem was in connection with the design of logging sleighs. Mr. W. E. Wakefield of the Forest Products Laboratories of the Federal Department of Mines and Resources conducted this research work and has published a valuable report (16). Both this work, and that of Mr. Klein are linked to similar studies carried out in Sweden, notably by Soderberg.

Although slow-moving sleighs are still invaluable for much winter transport, faster moving vehicles are an inevitable result of the application of automotive practice to this unusual field of transport. Many ingenious over-snow automobiles have been developed. One of the most successful is a Canadian machine, the « Auto-neige » designed and manufactured by Mons. Bombardier of Quebec. Over six hundred of these vehicles are successfully in regular use in Canada. This was the machine which provided the basic ideas for the Canadian Army's « Snowmobile » developed during the war years and given its first large scale public trial on « Exercise Musk-Ox ». This remarkable journey was made by a special Moving Force of the Canadian Army using twelve snowmobiles, supplied and assisted by R.C.A.F. planes. The Force left Churchill, Man., on February 14th, 1946 and reached Edmonton, Alta., on May 6th, after a journey of 3130 miles, penetrating far north of the Arctic Circle, most of which was made under most severe winter conditions (17). The entire Exercise provided much invaluable information about northern conditions, not the least important being that over-snow travel with automotive equipment was possible over a route which included conditions almost as bad as could be met with anywhere. The economics of this method of travel, however, have yet to be investigated. Enough is known of the operation of existing vehicles to make further improvements imperative before they can be widely adopted. The whole problem of « over-snow » travel is a challenging one; it seems reasonably certain that considerable advance is to be expected over present -day methods.

Importance of Fundamental Studies :

It is well realised that the foregoing brief review is essentially a listing of what might be called applied scientific or engineering problems. Were it not for one common feature the review might be inappropriate for such a meeting as this. All the problems involve water in one of its solid forms— snow and ice being the common terms used to describe the more common varieties of the frozen form of water. Empirical approaches can usefully be made to individual problems without any consideration being given to such a common link. When, however, a scientific approach is made to such problems, the common factor becomes of prime importance. Snow and ice as materials call for fundamental study, basic to any real understanding of the more applied problems which have been considered.

This view of the matter was adopted by the Associate Committee on Soil and Snow Mechanics of the National Research Council of Canada very early in its work

relating to over-snow travel. Constituted to deal with a special war problem, and including within its membership (as is the practice with N.R.C. Associate Committees) both civilian and military personnel, the Committee is now concerned with all the terrain of Canada with special reference to the operation thereon of tracked vehicles. Soil, « permafrost » and snow and ice are the materials under study through the Committee. From the outset of its work, their interrelation has been kept in view, as has the need for basic study of the materials in question as an essential foundation for more applied researches.

The Committee has been fortunate enough to have Mr. G. J. Klein as a member since its inception. In his work on areoplane skis, Mr. Klein had found the need for a basic study of snow as a material. He therefore brought to the Committee a background of useful preliminary investigation. Against this background and with his invaluable advice and the help of the Dominion Meteorological Service and other agencies, the Committee was able to arrange for the Snow Survey across Canada which is described by Mr. Klein in his companion paper (18). As is therein explained, the main object of the Survey was to provide actual knowledge as to the properties, and not merely the water equivalent, of the snow encountered at selected locations across the Dominion. Without this knowledge, no detailed snow researches could be carried out effectively.

A review of the literature which was naturally studied as a start to these Committee projects, showed that relatively little had been done on snow and ice as materials. As the most suggestive publications were those from Switzerland, personal contact was established with Swiss workers with the most helpful results. Dr. Haeffeli of the Zurich Polytechnique and Dr. Bucher of the Snow Research Institute at Davos, and members of their staffs, have not only been most courteous in the provision of information but their own work and their achievements have been a source of inspiration to the start of Canadian research into snow and ice.

As a further preliminary, the Committee arranged for the holding of a Conference of those Canadians experienced in snow work. This was held at Ottawa in September 1947; a record of the proceedings is available (19). The Conference was organised jointly by the Committee already named and the corresponding Associate Committee on Geodesy and Geophysics (the National Committee for Canada of the International Union of Geodesy and Geophysics). The hydrological aspects of snow and ice were recognised as an important part of the general picture, and that part which had received a good deal of public attention. It was hoped that the Conference would show the complete interrelation of the hydrological and the many other aspects of snow and ice study then appreciated. These hopes were more than fulfilled. Seventy experts attended the Conference and in the course of two days discussed well over one hundred different problems associated with snow and ice. A partial list of these accompanies this paper as an Appendix. Most striking was the repeated emphasis upon the necessity for a knowledge of the basic properties of snow and ice as materials. Sir Charles Wright, who was glaciologist to the Scott Terra Nova Expedition and who attended the Conference, was particularly emphatic on this point. This paper is, in effect, a digest of the proceedings of this important meeting.

The Conference decided that there should be a continuing Research Sub-committee, to be operated (in the first instance at least) jointly by the two Committees. The main purposes of this Sub-committee would be to correlate work being done on snow and ice in Canada and to promote the initiation and development of further work to the maximum extent possible. The Chairman of this Joint Sub-Committee (Mr. P. D. Baird) is present at this meeting of the International Commission, as are also Mr. Klein and the Chairmen of the two main committees mentioned (Dr. J.T. Wilson and the writer). This may be some evidence of the keen desire of Canadian workers to associate themselves with fellow workers in other lands and of their hope that the remarkably broad aspects of snow and ice research, which have been so briefly sketched in this paper, going far beyond the confines of hydrological interest, may commend themselves to those from other countries in which snow and ice are important features.

The Future :

There has been in recent years in Canada a remarkable awakening of interest in the northern part of the country. As an indication of this there may be mentioned the inauguration of the Arctic Institute of North America, with headquarters in Montreal, which is mobilising public interest and support for Arctic investiga-

tions (20). This public attention to northern problems carries with it an appreciation, at least of a general character, of the importance to the Dominion of snow and ice. It is, therefore singularly appropriate that circumstances have led to the renewed scientific activity in the same direction which is indicated in this paper. It can consequently be predicted with a reasonable degree of certainty that real advance in the fundamental study of snow and ice may be expected in Canada in the years immediately ahead.

This advance will be a cooperative effort on the part of all the agencies mentioned, directly and indirectly, on the foregoing notes. There will be the most complete pooling possible of information assembled by interested private companies, and by municipal, provincial and federal governmental agencies. The special interests of the Dominion Water and Power Bureau are being described to this Conference by its Controller, Mr. V. Meek. The companion paper by the Controller of the Dominion Meteorological Service, Mr. A. Thomson (21), shows clearly the vital concern of this Service with problems of snow and ice, especially in view of the extension of its services in the far north.

It is in concert with such sister organisations that the National Research Council is working and will continue to work in this field. Through its committee service the Council provides means for active cooperation between all interested agencies. Through its operating Divisions — of Physics, Mechanical Engineering and Building Research — it prosecutes active research into appropriate winter problems. The Division of Mechanical Engineering is investigating aeronautical icing and aircraft ski design; it will be investigating over-snow transport and snow clearing problems. The Division of Building Research, in turn, will be concerned with static problems of ice and snow and consequently of the fundamental properties of these materials. In conformity with Swiss practice, this work on « Snow Mechanics » will be carried on in close association with similar work in Soil Mechanics. Within the next few years it is hoped to establish snow research stations in the Canadian north and in the mountains of the continental divide. The National Research Council anticipates with pleasure the active assistance of Swiss workers in this work, this being further evidence of the desire on the part of Canadian authorities for international cooperation in this field of scientific interest such as is so fortunately the case in so many other branches of science. It is hoped that this Conference will result in real advance towards this most desirable goal.

APPENDIXE

List of Canadian Research Projects on Snow and Ice Suggested at the Ottawa Conference, 1947

I. *Fundamental Research :*

1. Establishment of observation and research stations.
2. Correlation and expansion of existing snow surveys, and standardization and distribution of data so obtained.
3. Maps of snow cover and its seasonal variation.
4. The physics of snow-melting, particularly as related to absorption of solar heat.
5. Knowledge of the properties of various types of ice, and conditions under which they are formed.
6. The importance of cyclic trends of solar radiation as affecting severity of weather.
7. Seasonal forecasting of snowfall.

II. *Hydrology and Meteorology :*

1. Correlation of snow accumulation and run-off, finding variables involved and their individual importance and effect.
2. Accurate prediction of run-off as affecting power development and flood conditions.
3. Influence of snow meltwater on ground water storage.

III. *Snow Control and Clearance :*

1. Location of existing types of snow fences and windbreaks for highest efficiency.
2. Development of more efficient means of control of snow drifting.
3. Improvement of snowplow design.

4. Development of more efficient means of snow and ice removal.
5. Development of non-injurious chemicals for road treatment.
6. Development of anti-slip treatment for roads and sidewalks.
7. Design of self-cleaning highways.

IV. *Oversnow Travel :*

1. Properties of snow as affecting skis, sleds, or toboggans, and effect of heating sliding surfaces.
2. Design of over-snow tracked vehicles.
3. Starting of tracked vehicles in cold weather.
4. Take-off and landing with aircraft skis.

V. *Agricultural Effects :*

1. Evaporation from snow surfaces as affecting snow storage.
2. Amount of run-off available for irrigation purposes.
3. Preservation of moisture in dry farming areas by drift prevention.
4. Run-off as affecting erosion and soil conservation.
5. Value of snow cover (a) as a fertilizer, (b) as protection against frost penetration, and (c) as promoting the slow absorption of water by soil.
6. The effect of infiltration of snow meltwater into frozen or semi-frozen soil.
7. The effect of permafrost on crops and plant growth.
8. Frost in problems of land clearing.
9. Snow mechanics as affecting the trafficability of farm vehicles.

VI. *General Ice Problems :*

1. Physical properties of ice, such as strength and friction.
2. Means of reducing or increasing such properties.
3. Conditions governing the formation of various types of ice, particularly frazil ice, and their critical nature.
4. Means of preventing the formation of ice, such as frazil ice or that against dams and other structures.
5. Pressure of large ice sheets on engineering structures.
6. Prevention of ice jams with consequent floods and structural damage.
7. Study of living glaciers.
8. Effect of rain on ice.
9. Rapid means of detecting thickness of sea ice.
10. Rapid means of estimating strength of sea ice.
11. Storage of ice for refrigeration purposes.

VII. *Ice Problems Affecting Navigation :*

1. Eliminating and controlling formation of ice on various parts of a ship.
2. Means of facilitating the removal of such ice after formation.
3. Means of predicting the close of the navigation season.
4. Means of lengthening the navigation season on the Great Lakes.
5. Means of clearing ship channels more effectively.
6. Movements of ice under wind, and controls for such movements.
7. Navigation in Arctic drift ice.
8. Forecasting of iceberg seasons.
9. Tracing the path of icebergs.
10. Design of ice-breaking vessels.

VIII. *De-Icing Problems :*

1. Data on super-cooled liquid particles in air, and their characteristics when encountered by moving objects.
2. Means of de-icing electrical conductors.
3. Elimination of « galloping » conductors.
4. Icing of aircraft propellers.
5. Icing of aircraft wings and fuselage on the ground.
6. Icing of aircraft in the air.

IX. *Permafrost :*

1. Identification and mapping of permafrost areas.
2. Determination of the nature of soils in which permafrost occurs.
3. Determination of the thermal conductivity of frozen ground.

4. Means of lowering the level of permafrost.
5. Foundation on permafrost.

X. Engineering :

1. Frost boils and frost heaving of roadbeds.
2. Depth of frost penetration as affecting the depth of laying pipe services.
3. The validity of the use of the number of degree-days of freezing as a factor in runway design.
4. Airborne machinery necessary for the construction of airports on rolled snow or ice-covered lakes.
5. Construction of ice roads for tractor trains.
6. Means of transporting and storing fuel in the far north.
7. Damage to wharves caused by the ice movements.

XI. Electromagnetic Effects :

1. Electrostatic charges on snowflakes.
2. Determination of snow and ice depth by electromagnetic waves.
3. Effect of blowing snow and ice crystal haze on ultra-high frequency radio waves, and on navigation and communication.
4. Detection of snow storms by electromagnetic waves.

XII. Comfort Problems :

1. Clothing in the far north.
2. Physiological and psychological effect of proper clothing and food.
3. Housing in the far north.

XIII. Miscellaneous :

1. Prediction of snow slides in mountainous areas.

REFERENCES

- (1) This statement appeared in the *Montreal Transcript* (a newspaper long defunct), in 1857; reprinted in *Maclean's Magazine*, Vol. 61, p. 60, July 1, 1948.
- (2) SHUTT, F. T., « Nitrogen Compounds in Rain and Snow », *Trans. Roy. Soc. Can.*, Sec. III, Vol. 19, pp. 1-10, 1925.
- MEEK, V., « *Glacial Observations in Canadian Cordillera* », Paper presented to Eighth General Assembly, I.U.G.G., Oslo, 1948.
- (4) GLOVER, T. R. and CALVIN, D. D., « A Corner of Empire », *Cambridge University Press*, 1937. (see particularly Chapter Six.)
- (5) PERUTZ, M. F., « Iceberg Aircraft Carrier », *Journal of Glaciology*, Vol. 1, pp. 95-105, 1948.
- (6) BARNES, H. T., « The Physical Constants of Ice », *Trans. Roy. Soc. Can.*, Sec. III, Vol. III, pp. 3-24, 1909.
- (7) BARNES, H. T., « Ice Engineering », Renouf Publishing Co., Montreal, 1928.
- (8) PRATLEY, P. L., « Collapse of Falls View Bridge », *Eng. Jour.*, Vol. 21, pp. 375-81, 1938.
- (9) MURPHY, J., « Ice Problems in Power Development », *Can. Eng.*, Vol. 58, pp. 607-611, 1930.
- (10) BROWN, E., and Clarke, G. C., « Ice Thrust in Connection with Hydro-Electric Plant Design with Special Reference to Plant at Island Falls on the Churchill River », *Eng. Jour.*, Vol. 15, pp. 18-25, 1932.
- (11) GERMAN, H. H., « M. V. Abigweit—Prince Edward Island Car Ferry », *Eng. Jour.*, Vol. 31, pp. 2-10, 1948
- (12) Rear Admiral H. F. JOHNSON, « Development of Ice-Breaking Vessels for the U.S. Coast Guard », *Trans. Soc. of Naval Architects and Marine Engineers*, Vol. 54, pp. 112-114, 1946.
- (13) KERRY, J. G. G., « Winter Temperature Cycle of St. Lawrence Waters », *Eng. Jour.*, Vol. 29, pp. 26-34, 1946 (discussion), *Eng. Jour.*, Vol. 29, pp. 521-35, 1946.
- (14) FINNIE, E. A., « Snow Drift Control by Highway Design », *Engineering Experimental Station Bulletin*, East Lansing, Mich., Vol. 15, No. 2, 1939.
- (15) KLEIN, G. J., « The Snow Characteristics of Aircraft Skis », Aeronautical Report AR-2, *National Research Council*, Ottawa, 1947.
- (16) WAKEFIELD, W. E., « The Efficiency of Logging Sleighs for Pulpwood Operations in Different Types of Terrain », Final Report, Project 107, Forest Products Laboratories, Department of Mines and Resources, Ottawa, 1938.
- (17) ROWLEY, G., « Exercice Musk-Ox », *Geographical Journal*, Vol. 109, pp. 175-185, 1947.
see also
BAIRD, P. D., « Exercice Musk-Ox », *Journal of Glaciology*, Vol. 1, p. 75, 1947.
see also
SANGSTER, A. G., « Some Mechanical Aspects of Exercice Musk-Ox », *Eng. Jour.*, Vol. 31, pp. 169-72, 1948.
- (18) KLEIN, G. J., « Canadian Survey of Physical Characteristics of Snow Covers », paper for International Commission on Snow and Glaciers, Oslo, 1948.
- (19) National Research Council, « Proceedings of 1947 Conference on Snow and Ice », *Technical Memorandum* N° 10 of Associate Committee on Soil and Snow Mechanics, Ottawa, October, 1947. (Copies available on application to the Secretary of the Committee, c/o N.R.C., Ottawa, Canada).
- (20) The Arctic Institute of North America, « A Program of Desirable Scientific Investigations in Arctic North America », *Bulletin*, N° 1, Montreal, March, 1946.
- (21) THOMSON, A., « Expansion of the Canadian Meteorological Service since 1939, Paper for International Union of Geodesy and Geophysics, Oslo, 1948.

15) CANADIAN SURVEY OF PHYSICAL CHARACTERISTICS OF SNOW-COVERS

by

G. J. KLEIN, Division of Mechanical Engineering

Summary

During the past two winters, the specific gravity, hardness and other physical characteristics of each layer in the snow-cover were measured once each week at a number of observation stations. The aim of the measurements was to obtain specific data on snow conditions in different areas which would be useful in many problems and which would form the basis for future snow studies. The method and instruments are described, and the results of the first winter's observations are presented.

Introduction

Large scale systematic snow observations have formerly been of three kinds : (a) meteorological measurements of snowfall, (b) hydrological snow surveys to determine the quantity of water stored as snow in river basins in order to predict the spring run-off, and (c) snow surveys for avalanche control as developed in Switzerland. Of these, only the latter have been concerned with the physical condition of the snow.

The properties of different forms of fallen snow vary so greatly that data on snow depth alone are quite inadequate for such problems as snow removal from highways, railways and airports; the development of over-snow motor vehicles and the resistance of sleigh runners and aircraft skis. The scarcity of snow-cover data appropriate to these and similar problems, some of which are of considerable economic importance, led the Associate Committee on Soil and Snow Mechanics of the National Research Council of Canada to conceive the idea of the survey herein described and to develop the technique and instruments which have now become the standard in Canada for obtaining specific data on fallen snow.

The survey has been in operation during the past two winters and was conducted by the Meteorological Service of the Department of Transport and the Associate Committee on Soil and Snow Mechanics of the National Research Council of Canada.

In the following description « snow » refers to a mass of snow rather than an individual grain or crystal.

Object of the survey

The object of the survey was to carry out periodic measurements of the physical characteristics of the snow-cover at a number of observation stations in order to obtain useful data applicable to a wide variety of winter problems, especially to those in which the condition of the snow is an important factor. The survey was also intended to provide a basis for future studies of the fundamental properties of the various forms of fallen snow.

Observation stations

Figure 1 shows the distribution of the observation stations in the 1947 survey.

A fair comparison of the results of several stations is difficult except when there is uniformity in the amount of shelter from sun and wind. Because intermediate degrees of shelter are not easily defined the survey was limited to two types of stations : *exposed stations* with a flat unprotected test area free from disturbances due to trees, buildings, etc., and *sheltered stations* with a test area exposed only to precipitation, e.g., a narrow strip along the south edge of a small clearing in



FIG. 1



FIG. 2

moderately to heavily wooded country. Only exposed stations were used in the 1947 survey.

The test area at Old Glory Mountain was adjacent to the Meteorological Station at the top of the mountain; all others were on or adjacent to airports.

At Churchill, completely exposed areas are unsuitable for snow measurements because of the prevalence of shifting bare patches which are produced by strong, steady winds. A test area slightly sheltered from the wind but exposed to the sun was therefore chosen at Churchill.

The principal topographical features of each station are given below.

- | | |
|-----------|--|
| Arnprior | — Very flat, open farm land with scattered wooded areas. |
| Churchill | — Very flat barren country with practically no trees. |
| Edmonton | — Elevation 2200 ft. — very flat, open farm land with scattered wooded areas. |
| Gander | — Flat, heavily wooded country. |
| Goose Bay | — Flat, heavily, wooded country with low brush in the immediate vicinity of the airport. |
| Malton | — Very flat, open farm land with scattered wooded areas. |
| Moosonee | — On the bank of a river in very flat barren country with some thinly wooded areas. |

- Old Glory Mt. — Elevation 7790 Ft. — surrounded by mountains of approximately the same elevation.
- Whitehorse — Elevation 2400 — on the bank of a river in a heavily wooded valley with mountains of 6,000 to 7,000 ft. elevation within 10 miles in most directions.
- Winnipeg — Very flat, open farm land with very few trees.
- The elevations which have been omitted in the above table were all less than 800 ft. above sea level.

Fundamental considerations

The physical properties of a material generally depend upon certain of its basic features such as composition and structure. Experience has shown that this is true for snow.

Snow is a very porous material. Taking wet snow as the general case, it may be considered to be a mixture of ice, air and water. The relative proportions and the physical properties of each constituent will therefore influence the physical properties of the snow. Since the properties of ice vary with temperature, temperature will be a contributing factor.

The size and shape of the grains which make up a mass of snow have a considerable influence upon its properties. New snow, because of its fragile crystals is structurally weaker than old snow which has grains of compact form, and the cohesion of wet snow depends upon grain size just as the cohesion of damp sand depends upon its grain size.

The snow grains often become bonded together during settling. This bonding can add considerably to the structural strength of the snow.

We may therefore consider the basic features upon which the physical properties of snow depend and which distinguish one form of snow from another to be : (a) temperature, (b) relative proportions of ice, air and water (c) average size and average shape of its grains, and (d) degree of bonding between the grains.

Temperature and average grain size can be measured without difficulty; the relative proportions of ice, air and water can be expressed in terms of specific gravity and percentage free water content, and average grain shape may be described with reference to a suitable chart such as the « Grain-Form Density Scale » adopted in the survey.

The most direct measure of the degree of bonding is the tensile strength of the snow. Tensile strength measurements, however, are rather inconvenient for routine use in the field. A more common measurement is that of snow hardness although it is perhaps a little less direct.

Any layer in a snow-cover can therefore be described by measurements of :

- (1) snow temperature,
- (2) specific gravity,
- (3) percentage free water content,
- (4) average grain size,
- (5) average grain shape, and
- (6) snow hardness.

Instruments

The instruments, which were designed especially for the survey, had a number of convenient features. The complete set, shown in Figure 2, weighed 16 lbs. including the carrying case.

Snow Sample Cutters

The cutters and knife shown in Figure 3 were used to obtain samples of 250 cc. volume. In determining the specific gravity of a snow layer, either two or four samples from the layer, i.e., 500 or 1,000 cc., were weighed at a time. The soft snow cutter, shown at the left, had a handle shaped to facilitate rotation about its axis in order to avoid compressing the snow. The removable cover plate was used only when samples tended to slide out of the cutter or when the specific gravity of a cohesionless layer was being determined. The hard snow cutter is shown at the right. The knife was used to trim the ends of the samples.



FIG. 3

FIG. 4

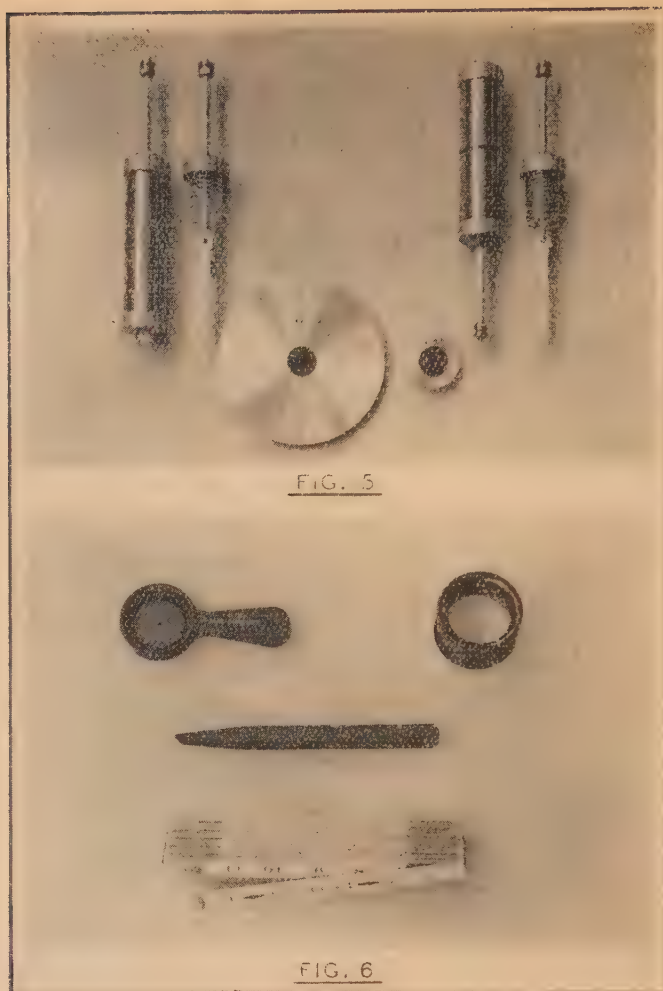
Balance

The beam type balance, shown in Figure 4, was ideal for specific gravity and percentage free water content measurements. It was more accurate than a spring balance and its auxiliary rider simplified free water content measurements. The auxiliary rider contained a spring clamp which could be released by pressing the button on the rider. The counterweight at the left could be raised or lowered to alter the sensitivity without disturbing the zero adjustment of the balance.

Hardness Gauge

Various instruments have been employed to measure snow hardness such as the Swiss «Kegelsonde» (1) and Nakaya's impact cone mounted on a pendulum (2). The gauge used in the survey was a spring balance type similar to one developed by Eugster and referred to by Seligman (3), and was chosen because it was very compact and covered an extremely wide range.

Two gauges are shown in Figure 5; the one at the left is the low hardness gauge and the one at the right is the high hardness gauge. The latter had a spring ten times as stiff as that of the former. The corresponding spring and graduated push rod are shown beside each gauge. The discs at the bottom of Figure 5 had areas of 100 and 10'cm² and were held in place on the end of the push rod by the friction



of a rubber washer cemented to the back of each disc. Two smaller discs having areas of 1.0 and 0.1 cm² were permanently attached to the end of the rod.

The gauges were generally used horizontally against the wall of a test trench. When used vertically, a small correction was made for the weight of the moving parts. The reading was obtained by slowly pressing the gauge against the snow and noting the value on the graduated scale at which the disc began to enter the snow. The hardness, in gms/cm², was obtained by multiplying the reading by the corresponding factor given below.

Disc Diameter cm	Disc Area cm ²	Multiplying Factor
11.28	100	1
3.57	10	10
1.13	1	100
0.36	0.1	1,000

The low hardness gauge was normally used with only the 100, 10 and 1 cm² discs (0 to 1,000 gms/cm²) and the high hardness gauge with only the 1 and 0.1 cm²

discs (1,000 to 100,000 gms/cm²). Readings to the nearest half division were considered sufficiently accurate.

Cup, Magnifying Glass and Spatula

These are shown in Figure 6 and were used for observing the size and shape of snow grains. The circles engraved on the bottom of the cup had radii which varied by one millimeter steps and formed a scale for observing grain size. The spatula was used to place and arrange the granules in the cup and to break up aggregates. A folding scale which was used for depth measurements is also shown in Figure 6.

Thermometers

Each set of instruments contained two 100 to—10° C mercury filled thermometers and two 50 to —100° C liquid filled thermometers. Four thermometers with a 100 to —50° C range would have been more convenient but were not commercially available. The thermometers were laboratory grade, 12 inches long.

Grain-form density scale

While there is an infinite variety of snow grain forms, the important feature is the degree of compactness of the shape of the grains. This feature has been given the name «grain-form density».

The scale of grain-form density used in the survey is shown in Figure 7, and although the classification is more or less arbitrary this does not in any way detract from its usefulness. It was intended to be merely a scale of the degree of compactness of grain shape and not a classification of the various forms of snowflakes and snow grains. The purpose was to provide a simple scale of grain shape which would indicate the texture of a snow layer when both average shape, as described by the appropriate class in the scale, and average size of its grains were given. The criterion of grain-form density was taken as the relative structural strength of the grains or crystals due to shape alone. Each class refers to many different shapes of grains but all shapes in any one class have approximately the same influence on the properties of a mass of snow. While the classes may not be uniformly spaced an attempt was made to arrange them in proper order. A description of each class is given below.

Class	Description
-------	-------------

- | | |
|----|--|
| A. | Very slender needles, and plane crystals with very slender rays and not more than three pairs of delicate branches per ray. This class is limited to new snow crystals of very slender proportions and open pattern. |
| B. | Plane crystals of new snow with many delicate branches, and very feathery hoar crystals. |
| C. | Plane crystals and needles of new snow having somewhat more substantial form than classes A and B, and partly settled snow grains of similar structural strength. |
| D. | Plane crystals and needles of new snow having still more substantial form than class C, and partly settled snow grains of similar structural strength. |
| E. | Hexagonal plate crystals of new snow or hoar and cup crystals of hoar. The plate crystals may or may not have various forms of small notches in their sides or small extensions at their corners. |
| F. | New snow crystals in the form of columns having a length to diameter ration of four or less, and partly settled snow grains of similar structural strength. |
| G. | Graupel or soft hail, i.e., snowflakes which have received a thick coating of rime during their fall to earth. Flakes with thin to moderately thick coatings belong to the class preceding G in which their shape including the rime deposit would place them. |
| H. | Grains of settled snow and old snow with crystal facets. The facets are produced by sublimation and their presence indicates that no recent melting has occurred. |
| J. | Grains of settled snow and old snow with no crystal facets. Grains which have lost their facets by abrasion during drifting or by melting |

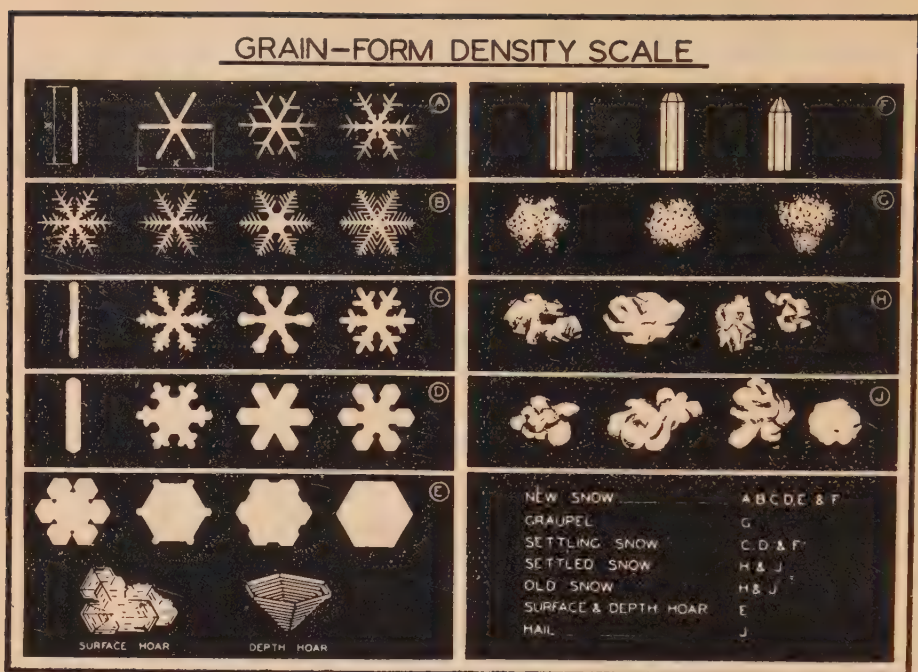


FIG. 7

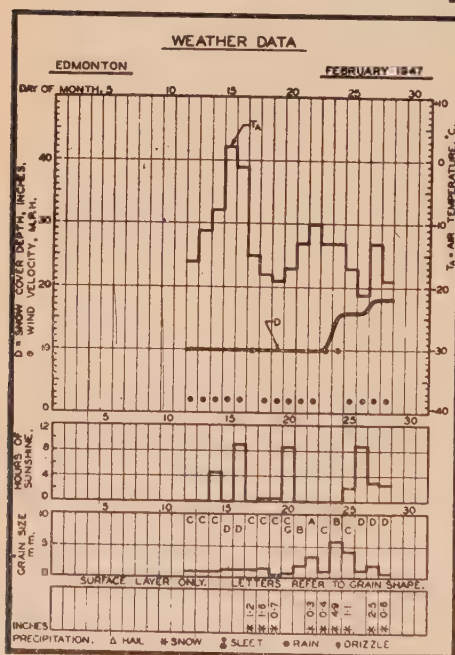


FIG. 8

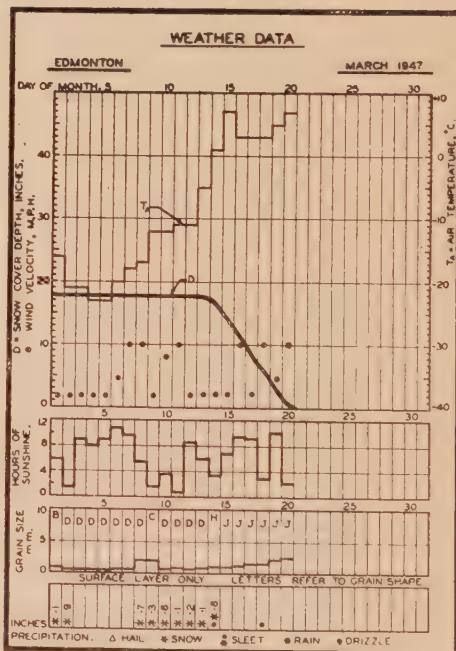


FIG. 9

which may or may not be followed by freezing, and hail belong to this class.

The above classes can usually be identified very easily although the facets of class H are sometimes difficult to recognize. A method which will serve to demonstrate the difference between classes H and J is to observe the amount of sparkle when direct sunlight falls on a sample of old snow. If facets are present, they will give a distinct sparkle effect; the fractures of broken grains are seldom flat and will not produce such clear cut reflections.

Procedure

The procedure followed at each station has been described in detail in Report MM-192 of the National Research Council of Canada (4). An outline is given below.

At each station a suitable area of about 500 sq. ft. was selected for the weekly and daily observations. A depth gauge in the form of a post marked off in inches was set up at one edge of the test area.

Weekly Observations

Once each week a new test trench, about 3 ft. $\times$ 3 ft. was dug down to the ground and the boundaries of the snow layers were located by examination of the section or by preliminary hardness tests. Air temperature, total depth of the snow-cover and the depths to the layer boundaries were measured, and snow temperature, hardness, average grain size, average grain shape and specific gravity were determined for each layer. Percentage free water content was only determined when the snow temperature was very near 0° C. and then for only three or four representative layers. Each trench was filled in on completion of the measurements.

Daily Observations

Observations of air temperature, total depth of the snow-cover, average size and average shape of the grains in the uppermost layer, wind speed, amount of sunshine, and the amount and kind of new precipitation were made each day. These were intended to establish continuity from one set of weekly measurements to the next.

Most of the daily and weekly observations were made at about 10:30 and almost all of the remainder at about 15:30 local time. The aim was to obtain approximately the daily mean value of air temperature.

Some observers reported wind at the time of the observation in miles per hour, while others used the Beaufort Scale or descriptive words. Further, the amount of sunshine was reported either in hours of sunshine for the previous day or as «overcast», «partly cloudy», «clear», etc. for the period since the last observation. This variation was no fault of the observers. Report MM-202 (5) was subsequently issued to ensure a greater degree of uniformity in the following winters' surveys. These amendments, however, were not in effect during the 1947 survey.

Description of the Measurements

Most of the measurements were quite simple and were made in the usual way. A few, however, require further explanation.

The temperature of a snow layer was always taken at the center of the layer thickness. Hardness, grain size and grain shape measurements were generally taken as average values for the layer.

Grain size was taken as the maximum dimension (X Figure 7) of a single grain.

Whenever new snow contained graupel, the average size of the graupel, as well as the average size and shape of all the other crystals were recorded, e.g., G 1.2, B 3.0; otherwise only the average size and average shape were recorded.

Percentage free water content was determined by a simple calorimetric method. From 300 to 400 gms of hot water, between 50 and 80° C, were placed in the balance bucket and its weight and temperature were measured. The balance was then set at zero and again balanced using only the auxiliary rider. A sufficient quantity of the wet snow was added to bring the temperature to between 5 and 15° C. The mixture was stirred to make sure that all ice particles were completely melted. The final temperature and the weight of snow which had been added were then determined. The percentage free water content was found by the use of nomograms for the following equation :

$$\text{F.W.C. \%} = 100 - \frac{5}{4} (T_1 - T_2) \frac{W}{S} - \frac{5}{4} T_2$$

where W = weight of water, gms.
 S = weight of snow, gms.
 T₁ = initial temperature of water, °C.
 T₂ = final temperature of water and melted snow, °C.

No corrections were made for the loss of heat to the air or for the thermal capacity of the bucket which was made of aluminium and weighed slightly less than 50 gms.

Wind speed, amount and kind of precipitation and the amount of sunshine were obtained from the records of the Meteorological Station adjacent to the test area.

The total depth of the snow-cover, the depths to the layer boundaries and the amount of precipitation were measured in inches and wind speed in miles per hour - the standard units in Canada. The characteristics of the snow, however, were either expressed nondimensionally or were measured in metric units to facilitate comparison with the comprehensive Swiss researches in snow mechanics.

Results of the 1947 survey

The results discussed here were obtained at exposed stations during the period from the middle of January to the middle of May 1947. They clearly demonstrate the value of surveys of this kind.

In order to facilitate analysis, the observations were plotted as graphs, examples of which are given in Figures 8 to 13. These graphs retain practically all details making it possible to examine the conditions under which certain changes occurred. They also reveal general trends and disclose features which might otherwise escape notice.

General Results

The results of all the stations taken together show several notable features of a general nature.

Air temperature and wind had a decided influence on the condition of the snow-cover, while the effects of sunshine were relatively small. The flow of heat from the earth appeared to cause some changes in the snow structure particularly in the lower half of the snow-cover. Isolation of the effects of any one factor, however, is not feasible because it is the combination of factors which is most important.

Wet snow conditions were surprisingly rare and were almost entirely confined to the short period of the spring thaw represented by the final steep slope of the snow-cover depth curves shown in figure 14. At Gander the main thawing period occurred late in February.

The maximum rate of ablation occurred at Goose Bay early in May when the snow-cover depth decreased 20'' in two days.

One inch below the surface the snow was harder than 100 gms/cm² in 70 % of all cases. Only rarely did the depth of moderately soft snow at the surface exceed 10 inches. Generally most of the snow-cover and frequently all of it had reached or passed the settled stage.

The most significant characteristic of a snow layer was hardness. Hardness, together with shape and size of grains would, in nearly all cases, give a fairly complete description of the layer.

Table I gives the range of specific gravity and hardness for the different kinds of snow layers.

There were relatively few cases of wet settling snow — often called « wet new snow » — and therefore the figures for this type should be regarded as only approximate.

The dividing line between settled and old snow was arbitrarily taken as 1 mm grain size.

Types 5 and 6 often contained depth hoar — in fact some layers were almost entirely composed of hoar crystals.

Snow Conditions at the Different Stations

The total depth of the snow-cover at the different stations has been plotted in Figure 14. It should be noted that at Goose Bay the snow-cover depth was much greater than at Gander which is only 400 miles south-east of Goose Bay. Gander and Goose Bay are about 30 and 130 miles inland respectively. The small amount of snow at Gander was due to fairly mild weather with the occasional rain.

Malton and Arnprior — about 200 miles apart — had quite different snow-cover depths. Temperatures and the frequency of precipitation were generally similar, but the amount of snow which fell at Arnprior was usually twice the amount which fell on the same day at Malton. The depth curves for the two stations have approximately the same general shape.

Air temperature and wind speed data for each station are presented in Table II. The values in the table refer to that part of the month for which the corresponding snow-cover depth curve in Figure 14 is shown as a solid line.

The snow conditions at the different stations are summarized below. The figures for snow hardness are all in gms/cm² and «settled snow» refers to snow which has reached or passed the settled stage, i.e., settled snow, old snow and loose granular snow.

Arnprior

Conditions near the surface varied from fairly soft snow to very hard crust. Classes H and J were about equally common and the snow layers were usually separated by thin ice sheets. The hardness of settled snow layers generally varied from 30 to 1,000. During the latter half of March there was a thick spring crust — hardness 8,000 — at the surface.

Churchill

Strong winds caused drifting and blowing snow on completely exposed areas. Most of the snow at the test area was deposited by the wind; only about one quarter was due to direct precipitation. The wind rapidly transformed new snow into settled snow which, since temperatures were very low, was always class H and had considerable sparkle in bright sunlight. There were never any ice sheets between snow layers which made it difficult to pick out the layers by eye.

The snow at Churchill and on the Barrens has a reputation of always being extremely hard but this was not entirely supported by the measurements — possibly because the test area was sheltered a bit from the wind. The hardness of settled layers generally varied from 100 to 1,000 with a few subsurface layers up to 4,000 hardness. On exposed areas the snow may have been as hard as that at Winnipeg where wind and temperature conditions were more or less similar. No extremely hard crusts which are formed by melting and freezing were found during the test period.

The size of settled snow grains increased with depth in much the same manner as shown for the lower half of the snow-cover in Figures 11, 12 and 13. This was more pronounced at Churchill than at any other station — possibly because of the absence of ice sheets between snow layers. The coarse layers near the ground, however, were generally fairly hard and dense and could not be described as «loose granular snow».

Edmonton

The entire snow-cover was relatively soft due, apparently, to the exceptionally low winds. Only one hardness reading of 600 was obtained — all others were 400 or less. Temperatures were fairly low and while crusts were reported these were seldom ice sheets. All settled snow layers prior to March 14th were class H.

Gander

Since there was very little snow at Gander, relatively few measurements were obtained. Wet spring snow with coarse grains, spring crust and ice were most common.

Goose Bay

Goose Bay had the deepest snow-cover of all the stations with the exception of the one on Old Glory Mountain. Temperatures were moderate and strong winds were common. From the start of the measurements until the end of April there was a layer, 12 to 15" thick, of loose granular snow — hardness 40 to 100 — near the ground, and throughout April there was a fairly thick crust — hardness 2,000

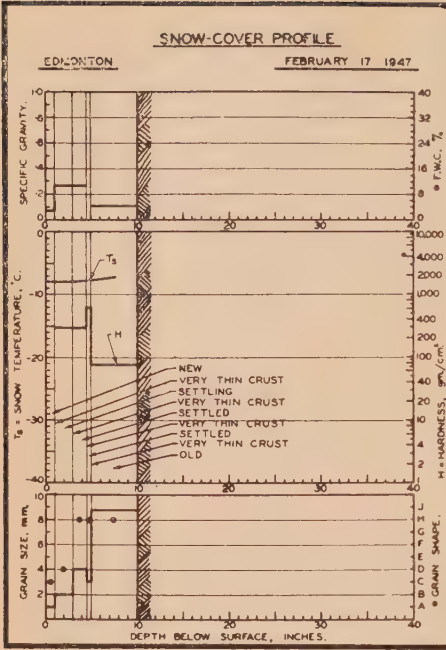


FIG.10

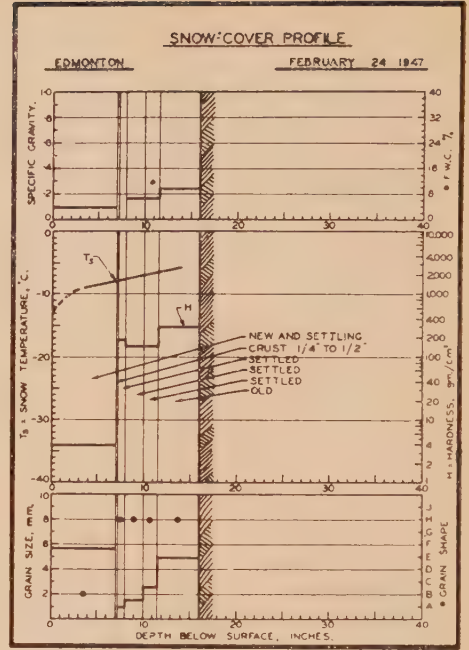


FIG.11

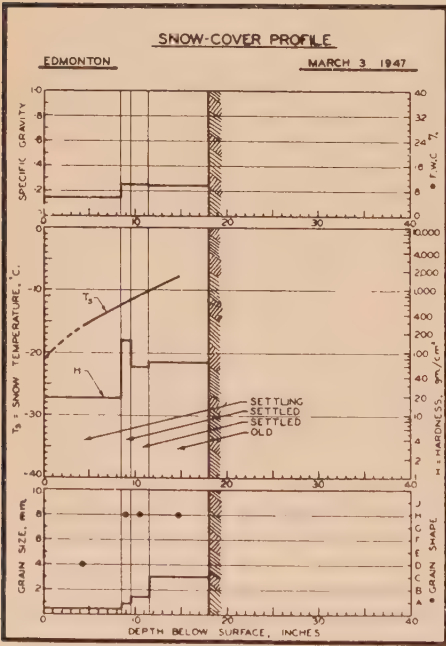


FIG.12

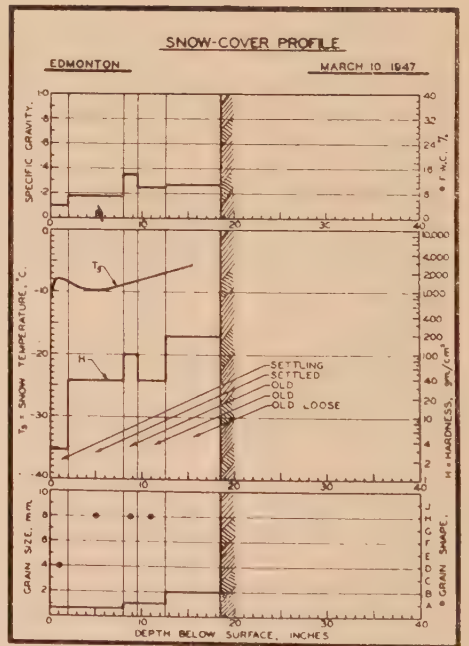


FIG.13

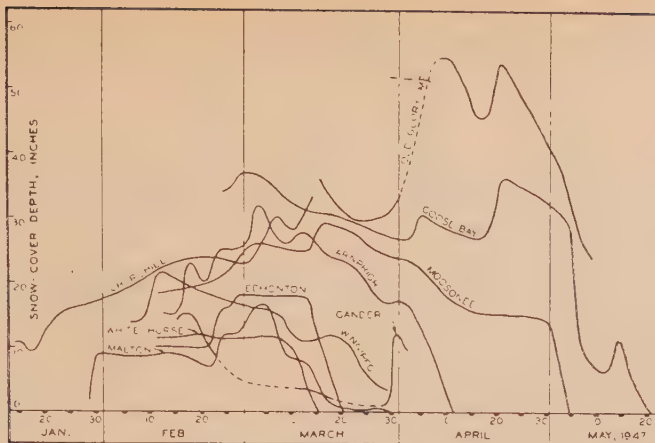


FIG. 14

to 4,000 — at the surface or immediately below freshly fallen snow. The hardness of all other settled snow layers was between 100 to 1,000. The high winds caused new snow to settle rapidly. With only a few exceptions the settled snow was of class J and thin ice sheets between layers were very common.

Malton

During the last two days of January a 4" layer of hail fell accompanied by rain. This layer became a crust — hardness above 10,000 — which was present in the snow-cover until the spring thaw began. There was also a solid ice layer on the ground which gradually increased to $3\frac{1}{2}$ " thickness at the beginning of the spring thaw. The hardness of all other settled snow layers was between 100 and 1,000. Settled snow was generally of class J and the layers were usually separated by thin ice sheets.

Moosonee

Up to the end of March there were no ice sheets between layers in the upper half of the snow-cover, but there were some in the lower half which may have been formed during the early part of the winter when the temperatures were not quite so low. Prior to the middle of April there was a layer of loose granular snow near the ground and all settled snow was class H. Thawing conditions began about the middle of April and resulted in very hard snow of class J throughout the snow-cover depth.

Old Glory Mountain

Relatively high temperatures caused new snow to settle very rapidly. Practically all layers had very coarse grains of class J and there were many ice sheets, some fairly thick, between the snow layers. The hardness of most layers was between 100 and 1,000. Toward the end of April the entire snow-cover became wet with a hardness range from 20 to 300. The maximum free water content during this period was about 15%.

Whitehorse

Settled snow layers were class J with hardness ranging from 20 to 1,000. Ground temperatures were lower than air temperatures indicating that the weather preceding the tests was appreciably colder than during the test period. The results obtained at Whitehorse, therefore, may not entirely represent snow conditions in that area.

Winnipeg

Steady winds and low temperatures were very common with the result that the entire snow-cover, except for an occasional thin layer of new snow at the surface, was hard wind packed snow of class H. Both hardness and specific gravity were definitely high for class H, the hardness varying from 200 to 8,000. The

snow grains at all depths and particularly near the surface were appreciably smaller than found elsewhere. A number of thin crusts, some of them ice, occurred between snow layers but they were not as common as at most other stations.

Conclusions

The results provide a great deal of general information about snow and the variety of forms in which it occurs on the ground, and clearly demonstrate that surveys of this kind yield data which can be of considerable assistance to both users and designers of winter equipment.

The 1947 survey gives a fairly clear picture of snow conditions on exposed areas across Canada. Subsequent measurements may alter this picture to some extent, but the consistency of the results indicates that the general outline will not change appreciably. If the survey is extended over about four winters it should provide a reasonably complete picture.

The results show that snow conditions on unsheltered areas, such as airports, are generally fairly hard and that the hardness of the snow on the Barrens, in spite of its reputation, is not greatly different from that on wind swept areas in most other parts of the country. In fact, the hardest form of snow, i.e., frozen spring crust, did not occur at Churchill during the test period nor for some time prior to the tests.

Although little would be gained by increasing the number of exposed stations, the value of the survey will be increased by the addition of an equal number of carefully selected stations sheltered from both wind and sun. It is hoped that the survey will be extended in this direction.

The instruments, grain-density scale and the general method proved to be entirely satisfactory. The system of weekly measurements supplemented by daily observations produced excellent results and made the best use of the observer's time and effort.

The method used in the survey provides a simple means for defining the physical characteristics of a snow-cover. It appears to fill a general need which has existed in many problems associated with snow.

Acknowledgements

The author wishes to express his sincere thanks to all who have contributed to the success of the survey. Thanks are due to the observers who faithfully carried out the observations, and to Dr Andrew Thomson, Controller of the Meteorological Service of the Department of Transport; Dr O.M. Solandt, Chairman of the Defence Research Board; Robert Legget, Chairman of the Associate Committee on Soil and Snow Mechanics, and J. H. Parkin, Director of the Division of Mechanical Engineering of the National Research Council of Canada for their valuable cooperation which is deeply appreciated.

BIBLIOGRAPHY

1. — *Schneemechanik mit Hinweisen auf die Erdbaumechanik*, R. HAEFELI. Eidgenossischen Technischen Hochschule in Zurich, 1939.
2. — *The Physics of Skiing, the Preliminary and General Survey*, U. NAKAYA et al. Journal of the Faculty of Science, Series II, Volume 1, No 9, Hokkaido Imperial University, 1936.
3. — *Snow Structure and Ski Fields*, G. SELIGMAN. MacMillan and Company, London, 1936.
4. — *Method of Measuring the Significant Characteristics of a Snow-Cover*, G.J. KLEIN. Report No MM-192 of the National Research Council of Canada, 1946.
5. — *Snow-Cover Measurements, Supplement to Report No MM-192*, G. J. KLEIN. Report No MM-202 of the National Research Council of Canada, 1947.

TABLE I

Type of Snow Layer	Grain Size mm		Specific Gravity		Usual Range		Hardness gms/cm ²	
	Min.	Max.	Min.	Max.	Min.	Max.	Min.	Max.
1. Dry New Snow	0.2	7	.048	.11	.07 — .10	20	0.5	1 — 10
2. Dry Settling Snow	0.2	5	.09	.22	.10 — .20	200	5	10 — 100
3. Wet Settling Snow	0.2	5	.10	.24	.15 — .20	100	20	20 — 100
<i>Class H</i>								
4. Dry Settled Snow	0.2	1	.20	.33	.23 — .30	8,000	25	80 — 800
5. Dry Old Snow (excluding 6)	1	7	.20	.34	.23 — .30	9,000	50	80 — 1,000
6. Loose Granular Snow	1	9	.11	.30	.18 — .28	200	15	20 — 100
<i>Class J</i>								
7. Dry Settled Snow	0.2	1	.20	.43	.25 — .35	6,000	25	100 — 6,000
8. Dry Old Snow	1	8	.20	.53	.25 — .45	20,000	50	100 — 20,000
9. Wet Old Snow	1	4	.28	.32	.35 — .50	500	20	50 — 500

TABLE II

Station	Average Air Temperature °C					Average Wind miles/hour				
	January	February	March	April	May	January	February	March	April	May
Arnprior		- 8.2	- 1.4	4.6			5	9	14	
Churchill	-27.3	-25.0	-13.5			15	14	9		
Edmonton		-13.1	- 9.0				3	5		
Gander ?		0.2	- 3.6							
Goose Bay		- 3.7	- 6.5					11	14	13
Malton	- 4.7	- 6.7	- 0.6	- 3.0	4.6		12	12		
Moosonee		-13.4	- 7.4	- 4.9	3.7			10	10	
Old Glory			- 2.3	0	6.5					
Whitehorse **		- 3.4	- 1.6							
Winnipeg		-15.3	- 4.6				15	13		

* Limited data due to scarcity of snow.
 ** See remarks on snow conditions at Whitehorse.

by

ROBERT SMITH-JOHANNSEN

General Electric Research Laboratory, Schenectady, New York

Snow Data

Snow data must be gathered to aid people in widely different fields of endeavor in their work. The data must be useful for snow surveys, and give the engineer information regarding water storage, run off, and behavior during a possible thaw. Logging operators, resort owners, and the military want to know how the snow cover facilitates or hampers various means of transportation. Highway and railway engineers are interested in knowing when avalanche danger is prevalent.

Snow Data

To accomplish all this effectively it is necessary that the data be readily transmitted, tabulated and understood.

Although one might say that there are an infinite variety of snow conditions, it is possible for all these purposes to classify them according to two basic conditions. They are the degree of modification of the original crystal form, and the degree of moisture saturation.

Unmodified snow may be defined as that which retains its crystalline identity. *Modified snow* has lost its original structure, perhaps even in the air on the way down to the earth. It consists of rounded particles of a great range in size, depending on their age and weathering conditions. *Dry Snow* is such that no liquid moisture is in contact with it, and consequently must be below freezing. Because the concentration of water vapor in the air around the particles is less the colder it is, the snow may be considered drier the lower the temperature. Dry snow usually tends to flow freely.

Wet snow is a mixture of snow and water. Its properties depend on the degree of modification of the snow particles, and what degree of saturation with water exists in it.

By dividing the chart into four areas representing these four main snow types, it is then possible to describe a definite snow condition in a graphical manner. This would then take into account the *degree* of modification and moisture saturation of the snow. Furthermore, this information could be contained in a combination letter and number code as is often employed in locating positions on maps.

Still more information may be given by the addition of a number denoting the thickness of the snow layer in question.

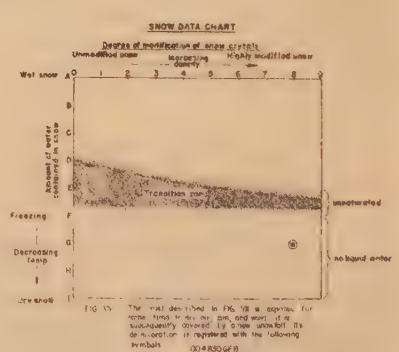
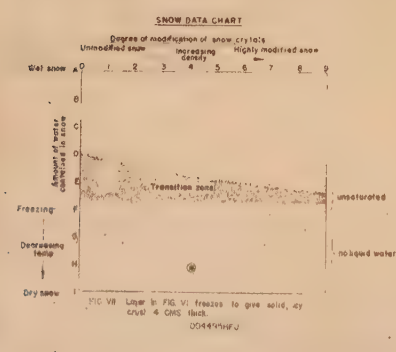
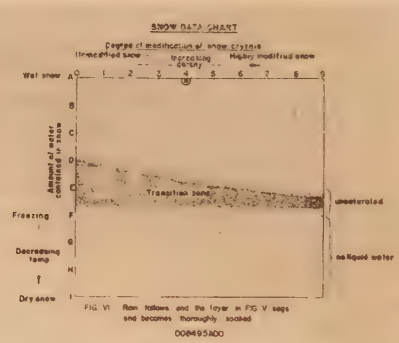
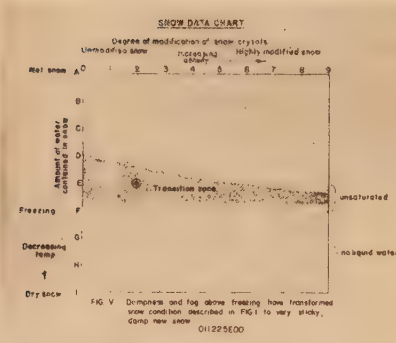
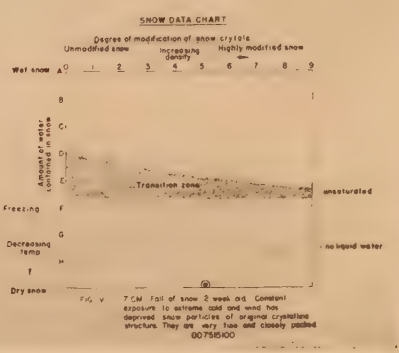
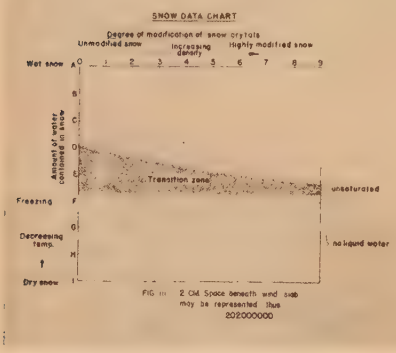
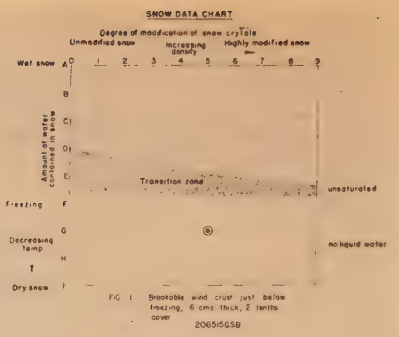
The symbols U and B indicate unbreakable and breakable crust respectively, the criterion being whether or not it will support a specific loading without breaking. Without this symbol it would be understood that the snow was loose.

The Symbols S and F indicate crust resulting from sublimation and freezing respectively. If the cover is spotty this will be shown by a figure denoting percent coverage.

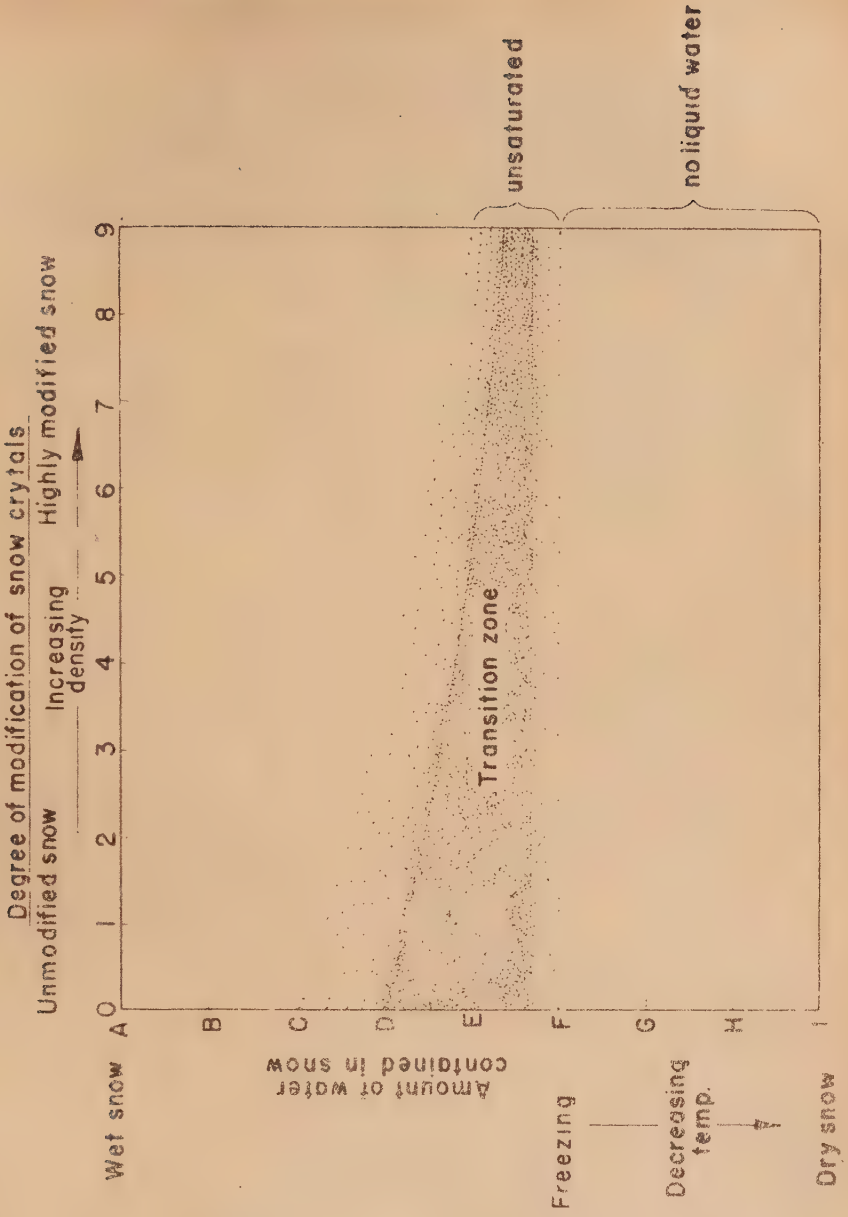
Snow Data Chart

Snow Data Chart :

The chart area is bounded horizontally by a scale marked from 0 to 9 representing the degree of modification of the snow crystals, and vertically by a moisture content scale ranging from A to I. Although the density of the snow layer generally increases with the modification of the snow particles, the two properties have such widely different effects, that it would be unwise to compromise one estimate with a consideration of the other. Therefore, density is not directly considered when assigning a specific snow condition, its position on the chart.



SNOW DATA CHART



The position of the point on the chart thus describes the nature of the snow condition.

For example, a specific condition might be described as IF or 3C etc.

Snow Data Symbols :

Other information is necessary, however, to complete the description of the snow cover. It is necessary to know the thickness of the layer in question and if it does not cover 100 % it is important to know approximately the percent of coverage. If the density of the snow can be measured it should be reported, and if there is a crust, it should be described as a sublimation crust or a crust due to freezing. It is also desirable to know also if the crust is breakable or unbreakable.

With the following code, I believe, the pertinent information about any snow layer, could be recorded.

Snow Data Code

1st Symbol	Per cent cover	(tenths)	One figure
2nd Symbol	Thickness of layer	(cms.)	Two figures
3rd Symbol	Degree of modification	(0-9)	One figure
4th Symbol	Density	(hundredths)	Two figures
5th Symbol	Moisture content	(A-L)	One figure
6th Symbol	Kind of crust	(S or F)	One figure
7th Symbol	Strength of crust	(U or B)	One figure
			0 indicates no crust

THE DEVELOPMENT OF A SNOW CLASSIFICATION SYSTEM

by

VINCENT J. SCHAEFER

General Electric Laboratory, Schenectady, New York

With the advent of air travel through polar regions, the necessity of the movement of mechanized equipment over snow fields, the increased importance of more accurate snow surveys for irrigation, power, and flood forecasts, the expanding use of snow surveys in the great mountain systems of the Himalayas and the Andes, and a rapid climb in the popularity of winter sports, especially skiing — these and other reasons emphasize the need for a simple, concise, practical and useful method for classifying specific snow conditions in such a way that no ambiguity remains when a specific snow condition is described.

To be satisfactory the method must be very simple; it must be easily understood, inclusive enough to become a universal, international system flexible enough to cope with the many special conditions which occur in the field, and yet, so easy to use that it tends to become the logical way to record technical observations of snow storms, snow conditions, and the snow pack.

Among the methods known to the writer, three shall be briefly discussed since they are typical of approaches which may be made to solve the problem.

Classification of Frozen Precipitation

When detailed observations are made of the various types of precipitation which occur during snow and sleet storms, it becomes apparent that definite forms of crystals are associated with different kinds of storms. Although there is an almost infinite variation in the fine structure of individual snow particles, it is possible to classify such precipitation into nine simple, basic forms. A number of such classifications have been proposed in the past, but in all cases known to the writer, the methods are either over simplified or too complicated for practical use in field studies.

The classification chart shown as Fig. 1 has been in active field use since 1943 in conducting studies of snow storms in the lowlands, mountains, and from airplanes in various parts of the United States and Canada. All precipitation forms observed in these studies could be assigned to basic forms listed in the chart.

A method of describing, in a standardized way, the physical properties of snow on the ground has been proposed by Mr. R. E. Lundquist as « An Outline of Descriptive and Quantitative Classifications of Various Conditions Persisting in Snow and Ice Deposits ». A careful study of this proposed method shows that most of the conditions encountered in natural snow and ice deposits, whether new or old, unmodified or highly modified, have been considered.

I believe that in the classification of « fresh accumulations », the differentiation implied by using the terms « snow grains » and « snow pellets » is not necessary since the formative process is essentially the same and the subsequent modification probably proceeds in an identical manner.

The inclusion of hail in a consideration of factors having an important effect on snow accumulation seems hardly necessary since it is of rare occurrence in most parts of the world during the snow storm period.

The inclusion of rime is also questioned for while it is very important on windy mountains commonly having supercooled orographic clouds, the vast majority of the snow cover in the world would not include such deposits. It must be admitted, however, that the snow fields which persist late in the season and thus are quite important would tend to have a certain amount of build-up from this source.

It is felt by the writer that the classification should differentiate more sharply between snow crystals which are basically of a solid compact type (such as sleet and the hexagonal plates, hexagonal columns, capped columns, and asymmetrical crystals) and those which tend to be less « icy » such as stellars, needles, powder, and graupel and any of the others which are rimed as they fall through supercooled clouds.

This situation could probably be remedied by using two classifications. One would be of a type like that proposed by the writer which includes only the solid precipitation common to snow storms, and the second would include the effects and processes related to the snow surface. For example :

1. Deposited Rime
2. Glaze
3. Unbreakable Crust
4. Breakable Crust
5. Frost
6. Hoar
7. Blowing Snow
8. Drifting Snow
9. Wind Pack

The fourth procedure, that of Robert Smith-Johannsen of the General Electric Research Laboratory in Schenectady, New York, includes several unique features, the outstanding one being a simple and, what appears to be, a practical graphical method for depicting the properties of the snow cover. A simple code system is used which greatly simplifies the transmittal of pertinent information from field stations to a central quarter and has the added value that it serves in lieu of a universal language.

The most attractive feature if this method is that it readily permits a visualization of the snow properties and the relative relationship of different kinds of snow when two or more types occur in the same deposit.

Conclusion

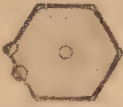
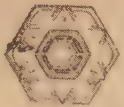
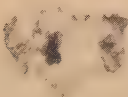
A combination and blending of these several methods and any other which may be proposed should be attempted by a working committee set up by the Commission so that within the next year or two a specific method may be ready for further critical use by snow observers. Before final adoption, the selected method should be given an intensive trial for at least a year at the various snow laboratories throughout the world.

TYPES OF SOLID PRECIPITATION

(see illustration)

1. **Stellar Crystals** : Six rayed, starlike forms occurring as single crystals or in clumps. Crystals vary in a wide range of form and size. Clumps common, often appearing as cottony flakes sometimes two inches in diameter. Size $1/32''$ — $1/2''$; clumps common.
2. **Graupel** : Stellar crystals which are covered with frozen cloud particles forming a rime deposit. This results in pellets of soft snow, which range from hexagonal to rounded forms. Size $1/16''$ — $1/4''$; rarely clumped.
3. **Hexagonal Plates** : Thin, hexagonal, solid or semi-solid plates often containing internal structure due to air inclusions. The smallest form of this crystal and the stellar form is known as Diamond Dust. Size $1/1000''$ — $3/16''$; rarely clumped.
4. **Hexagonal Columns** : Transparent generally flat-ended, sometimes pointed hexagonal prisms mostly of clear ice, many with air inclusions. Often a group of columns grow from a common source radiating in several directions. The former type produce 22° and 46° halos around the sun. Size $1/64''$ — $1/8''$; sometimes clumped.
5. **Capped Hex. Columns** : Similar to (4) with the exception that both ends terminate as expanded hexagonal plates. A hexagonal plate sometimes is positioned midway between the two plates. Many queer forms occur. Size $1/64''$ — $3/16''$; sometimes clumped.
6. **Ice Needles** : Long slender shafts with hexagonal cross section often terminating with sharp points. Needles occasionally grow in masses in random directions. Size $1/64''$ — $3/8''$; often clumped.
7. **Asymmetrical Crystals** : Angular crystals without a symmetrical outline occurring as simple or compound crystals. Size $1/64''$ — $1/8''$; often clumped.

8. **Powder Snow** : Dry bits of snow without angular form, often of irregular shape. Size $1/64''$ — $1/8''$; rarely clumped.
9. **Sleet** : Pellets of translucent to transparent ice often bearing bumps or other protuberances. They are frozen rain drops but are not always round. Size $1/32''$ — $1/4''$; very rarely clumped.

TYPES OF FROZEN PRECIPITATION			
CODE	FORM		TYPE
1			STELLAR CRYSTALS
2			GRAUPEL
3			HEXAGONAL PLATES
4			HEXAGONAL COLUMNS
5			CAPPED HEX. COLUMNS
6			ICE NEEDLES
7			ASYM. CRYSTALS
8			POWDER SNOW
9			SLEET

ABSTRACT OF FILM « SNOW HARVEST »

(16 m. m. color and sound)

by

Soil Conservation Service, U. S. D. A.

This 25-minute film shows in brilliant color winter scenes of snow covered mountains of the Western United States. It illustrates authentic travel adventures of engineers who measure mountain snow for the purpose of forecasting summer water supply to be available for irrigation, hydro-power generation and municipal use.

The technique of snow surveying is described. The organizational set-up for accomplishing the job is pictured.

There is shown also the system of reporting the results to water users and the economic uses served by snow surveys.

R. A. WORK

7-1-48

Medford, Oregon

U. S. Soil Conservation Service

NOTE: — This film must be screened using a 16 mm. sound projector. The running commentary is by Dr Vincent J. Schaefer of the General Electric Research Laboratory, Schenectady, New York.

THE PRODUCTION OF ICE CRYSTALS IN A SUPERCOOLED CLOUD IN THE LABORATORY

This short film shows two of the many experiments which may be conducted in an open top cold chamber. The first experiment shows the formation at a temperature of -15°C of a supercooled cloud by condensation of the breath. This is followed by seeding with tiny fragments of dry ice scratched from a piece of it held in the warm air above the chamber. The tiny fragments falling through the supercooled cloud leave a trail of millions of ice crystals.

The second experiment shows a similar result achieved by introducing a needle cooled below -39°C (the critical temperature) and passing it through a supercooled cloud.

The movie was produced in the laboratory of Dr. Vincent J. Schaefer and photography was by the chief cameraman of the American Museum of Natural History

18) THE DEPENDENCE ON ALTITUDE OF THE SNOW COVER IN THE ALPS

by

W. MÖRIKOFER

(Physical-meteorological Observatory, Davos, Switzerland)

I. Introduction

The problems of the snow cover and of its climatological conditions have met but recently an increasing interest. This interest is certainly founded in the knowledge that the problems of snow cover and of its variations are important not only for climatology itself, but also for numerous applications to great fundamental as well as practical tasks. In this regard snow research is interested in the physical properties of the snow cover as, for example, thermal conditions, reflection, transmission, absorption and emission of radiation, the specific weight, the transmissibility for air and water. The meteorologist deals with the correlation between weather influences and snowfall, depth and properties of snow cover. Avalanche engineering, a combination of engineering and natural research, which has been developed especially in Switzerland, is interested in depth and mass of snow cover, apart from all the questions of its cohesion and solidity. For hydrology and electrical economy the snow deposits of the winter represent a reserve of accumulated water for spring; to know its quantity is a fundamental problem for hydrology.

Let us mention some other fields where snow research is of more practical signification. For transport in the mountains the snow cover is of great importance; generally the snow cover is an obstacle to transport, sometime it can, however, facilitate traffic with remote regions. Even armies in mountainous regions are most dependent on snow cover, the reason why the Swiss army has recently developed a special avalanche service. Finally I may mention winter sports, in connection with which it is of interest to be informed of the depth of the snow cover to be expected and of the period during which the necessary minimum depth is assured (climatology of winter sports).

In spite of this importance of the problems of snow cover both for general aspects and for practical applications, our knowledge of the climatological properties of snow cover is very modest. There exist but few systematical investigations on the depth or on the duration of snow cover, because long series of snow cover observations are very rare; mainly in Austria a good work has been done about these problems. In Switzerland there exist very extensive records of snow cover observations which, partly, began 40 or 50 years ago. These observations have not been obtained at the ordinary stations of the meteorological network, but at the railway stations of the Rhaetian Railway.

At the Davos Observatory we have elaborated, together with O. Eckel (1) (2), G. Perl (3), F. Prohaska (4) and other collaborators, different problems based on these records, comprising almost 200.000 single observations. In the present report it is proposed to give a short review of some selected problems suggested by the records, especially as regards the dependence of snow cover on altitude. The regional properties will not be discussed in detail, but some principal aspects which may be of general interest.

First of all the regional conditions may, however, be described. The Grisons, the region served by the Rhaetian Railway, are the great easternmost canton of Switzerland, situated on the Central Alps. The country extends from 500 to 4000 m above sea-level, the snow measuring stations from 500 (Landquart) to 1800 m above sea-level (Upper Engadine). The country covered by our study comprises various valleys in the northern and central Grisons; these extend from the northern parts

(1) O. ECKEL, Ueber die Schneeverhältnisse von Davos. Jahresber. Naturforsch. Ges. Graubündens 75 109 (1938).

(2) O. ECKEL, Ueber die Schwankungen der Schneeverhältnisse in Davos. Meteorol. Z. 55, 359 (1938).

(3) W. MÖRIKOFER und G. PERL, Ueber die Abhängigkeit der Schneeverhältnisse von der Meereshöhe in Graubünden. Verhandl. Schweiz. Naturforsch. Ges. Sils 1944, 103.

(4) F. PROHASKA, Wetterlagen bei grossen Schneefällen in Graubünden. Verhandl. Schweiz. Naturforsch. Ges. Locarno 1940, 123; Jahresber. Naturforsch. Ges. Graubündens 78, 175 (1943).

of the Alps, which are quite exposed to moist winds, into the elevated central parts of the Rhaetian highlands, fairly protected by the outer mountain-chains from wind and precipitations.

The observations have been taken at 70 stations on the Rhaetian Railway; for the following study 21 of the most important and the most reliable stations have been selected, the observations extending over 30 years, 1913-1943. For several problems, where a long period is not required, the results of 13 years of observations at the new Federal Institute for Snow and Avalanche Research at the Weissfluhjoch above Davos, at an altitude of 2540 m, have been added.

Five valleys have been selected for this study, namely: the Prätigau, the Landwasser or Davos Valley, the Albula Valley, the Engadine, the Fore Rhine Valley.

A list of the stations is given in Table 1; the altitude of each and the abbreviations in the pictures are added.

Table 1. — Stations with snow cover measurements.

Valley	Station	Altitude, m/s.l.
Prätigau	Landquart	523
	Schiers	657
	Küblis	813
	Klosters-Platz	1194
	Cavadürli	1355
	Laret	1525
	Davos-Wolfgang	1629
Landwasser Valley (Davos Valley)	Filisur	1084
	Glaris	1457
	Davos-Platz	1543
	Davos-Wolfgang	1629
	Weissfluhjoch	2540
Albula Valley	Filisur	1084
	Bergün	1376
	Preda	1792
Engadine	Schuls	1290
	Zernez	1474
	Zuoz	1695
	Bervers	1714
	St. Moritz	1778
Fore Rhine Valley	Landquart	523
	Ilanz	702
	Truns	855
	Disentis	1133

Because all these stations are located near the railway-line they are situated in a comparable situation, in the main in the flat bottom of the valleys. On the other hand two defects of these records must be mentioned here: The purpose of the observations has been purely practical. The railway management is not interested in ascertaining the depth of the snow cover when it is low, and so the observations are not very exact and reliable at the beginning and at the end of winter, so that the duration of a snow cover of less than 20 cm cannot be studied with reliability. For the same reasons, only the total depth of the snow has been recorded; the depth of fresh snow can be estimated only from the increase in the total depth, taking into consideration the settling of the snow that has taken place. Because in this way exact figures cannot be obtained, any conclusions on fresh snow are not taken here.

The following calculations have been made from the records:

a) mean figures of the depth of the snow cover in the various parts of the Grisons for December to March, 1913-1943, these being the only months with reliable observations;

b) maximum figures of the depth of snow cover for the same places and the same period ;

d) mean figures for the duration, the beginning and the end of a snow cover of a depth of at least 20 cm.

II. Mean depth of the snow cover

Beginning with the average figures of the snow cover, we resume the 30 years' means for every of the four winter months December to March and their winter means for all stations of Table 1 in Table 2. Some examples of these results are illustrated by Fig. 1, 2 and 3, the dependence on altitude of the depth of snow cover for December and March for the five examined valleys and the monthly variation from December to March especially for the Prätigau, the valley with the greatest depths of snow cover. The small circles in the figures represent the stations, each of the lines a valley (respectively a month).

Table 2. — Monthly means of the depth of snow cover in cm (1913-1943) for December to March and for the winter.

Valleys and stations	Altitude m	Mean depth of snow cover				Winter Decemb.-March
		December	January	February	March	
<i>Prätigau</i>						
Landquart	523	8	15	15	4	10
Schiers	657	17	35	41	21	28
Küblis	813	23	43	51	26	36
Klosters	1194	38	66	80	62	61
Cavadürli	1355	52	80	99	89	80
Laret	1525	57	98	118	118	98
Wolfgang	1629	46	82	100	97	81
<i>Davos Valley</i>						
Filisur	1084	15	31	30	13	22
Glaris	1457	32	64	74	63	58
Davos-Platz	1543	35	65	77	67	61
Wolfgang	1629	46	82	100	97	81
Weissfluhjoch	2540	99	134	178	205	154
<i>Albula Valley</i>						
Filisur	1084	15	31	30	13	22
Bergün	1374	24	44	48	34	37
Preda	1792	37	61	71	72	60
<i>Engadine</i>						
Schuls	1290	24	43	47	32	36
Zernez	1474	25	46	51	41	41
Zuoz	1695	34	54	60	58	52
Bervers	1714	35	52	60	58	51
St. Moritz	1778	37	57	64	61	55
<i>Fore Rhine Valley</i>						
Landquart	523	8	15	15	4	10
Ilanz	702	15	33	38	17	26
Truns	855	25	50	55	29	40
Disentis	1133	26	48	52	37	41

The various results concerning the mean depth of snow cover can be summarized as follows :

1) The depth of snow cover shows a very regular increase with altitude for all examined valleys. Consequently, for any valley the snow depth may be interpolated for a locality between two stations of observation as a function of altitude.

2) The absolute depth of snow cover and the gradient of increase are different

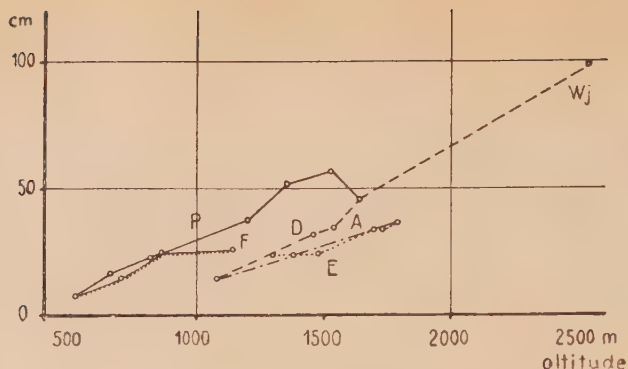


FIG. 1. Mean depth of snow cover in December, in different valleys of Grisons, as function of altitude. P : Prätigau, D : Davos Valley, A : Albula Valley, E : Engadine, F : Fore Rhine Valley, Weissfluhjoch.

for the various valleys, absolutely and with regard to the altitude. The snow cover in the exterior valleys, which are more exposed to wind and precipitation, is much deeper than in the central valleys, taken for the same altitude; the same rule is valid for the gradient of increase.

3) The curves for the different months show an analogous aspect, but their gradient is increasing during the winter. So, the variations of the snow cover are different at different altitudes. This is a consequence of the fact that, after increasing of the depth of snow cover in December and January throughout the whole region, for greater altitudes the depth of snow cover is increasing until March, whilst for lower stations it is decreasing from February (see Fig. 3). Consequently the gradient of the increase of the depth of snow cover with an increase of 100 m for the altitude is approximately the double in February and March than in December.

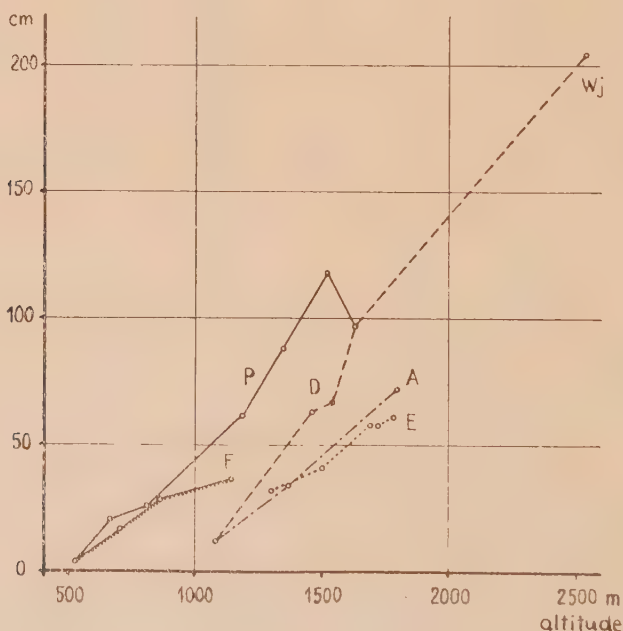


FIG. 2. Mean depth of snow cover in March, in different valleys of Grisons, as function of altitude.

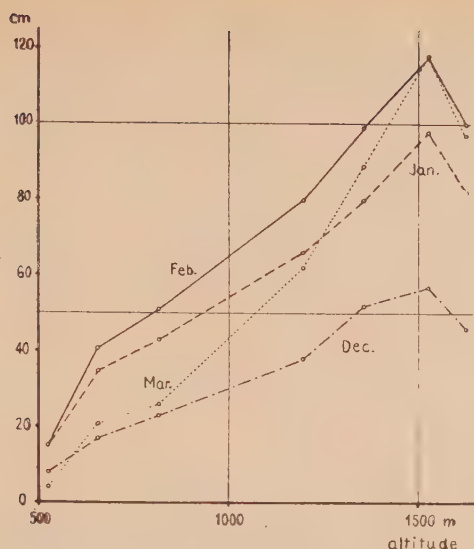


FIG. 3. Monthly curves of the mean depth of snow cover in the Prätigau, from December to March.

4) The interior valleys of the Grisons, especially the Engadine, are so poor in snow that at an altitude of 1800 m they have the same depth of snow cover as the exterior valleys at 1000 m.

5) The monthly variations of the depth of snow cover in the Prätigau, as shown by Fig. 3, can be summarized as follows : The depth of snow cover is, of course, lowest in December. It then increases during January and February; but in February the increase is considerable only in the higher regions, where no melting at all takes place, since in the lower part of the valley the increase is only small, due to occasional melting of the snow. This condition is accentuated in March, as in the higher parts of the Prätigau the depth of snow cover remains unchanged from February to March, whilst it is very much decreasing in the lower regions of this valley.

6) The curves which have been found by V. Conrad and O. Kubitschek (5) for the Austrian Alps show approximately the same figures for a given altitude as our curve for Prätigau, the Grisons valley with the deepest snow cover.

7) F. Steinhauser (6) has pointed out that in the Eastern Alps the effect of the upper limit of the temperature inversion at an altitude of 900 to 1100 m can be recognized with a discontinuity in the altitude curves of the depth of snow cover. Slight traces of this relation can be found also in our Grisons observations (see Küblis in Fig. 3); but it cannot be very pronounced because in this part of Switzerland the wide-spread stratus of fog above the ground in winter is quite seldom and not so frequent as in the northern Swiss Plane.

8) A very instructive example for the effect of wind protection by mountain-chains is given by a comparison of the simultaneous snow depths in the Prätigau and the Davos Valley, two valleys which join by a pass at Wolfgang (see Fig. 1 and 2). Prätigau, the valley which is more exposed to wind and precipitation has always a greater depth of snow cover than the Davos Valley, which is situated behind another mountain-chain; the difference of depth of snow cover between the two valleys, as compared between Laret and Davos-Platz (Table 2) or simply for an altitude of 1500 m (Table 3), increases from December with 22 cm (by 10 cm a month) to 50 cm in March. A further detail may be of interest : The steep ascent from Klosters to Laret produces a very high gradient of the increase of the depth of snow cover, even the greatest gradient of our observational data. Though

(5) V. CONRAD and O. KUBITSCHKEK, Beiträge zur Kenntnis der Schneedeckenverhältnisse : Die Veränderlichkeit und Mächtigkeit der Schneedecke in verschiedenen Seehöhen. Gerl. Beitr. Geophys. 51, 100 (1937).

(6) F. STEINHAUSER, Die Schneehöhen in den Ostalpen und die Bedeutung der winterlichen Temperaturinversion. Arch. Met. Geoph. Biokl. Ser. B. 1, 63 (1948).

Wolfgang in the same valley is situated 100 m higher than Laret, the depth of snow cover here is considerably diminished, because the valley there is turned, so that Wolfgang is situated behind another mountain-rigde.

In table 3 some numerical figures are given for the mean depth of snow cover at an altitude of 1500 m and for the mean increase with altitude, both for the different valleys of the Grisons and for the four months and the winter.

Table 3. — *Depth of snow cover in the Grisons.*

	Depth of snow cover at 1500 m					Mean increase in cm /100 m				
	Dec.	Jan.	Feb.	Mar.	Winter	Dec.	Jan.	Feb.	Mar.	Winter
Prätigau	56	94	115	112	94	5	7	9	11	8
Davos Valley	34	65	75	65	59	6	7	11	13	9
Albula Valley	28	49	55	45	43	3	4	6	8	5
Engadine	26	47	52	44	42	3	3	4	6	4
Fore Rhine Valley						3	3	4	5	4

By resuming the general results we find in the Alps a great increase of the depth of snow cover as altitude increases, a result which is not surprising at all. Three influences may be mentioned as cause of this increase with height :

1) the general increase of the amount of precipitation with altitude, which has been established more or less distinctly for all mountains,

2) the increasing percentage of snow as compared with rainfall in the precipitation,

3) the diminution of melting velocity as altitude increases, which is a consequence of the decrease in temperature; this third influence is especially the reason why the mean gradient of snow cover increases towards spring, because then the melting is advanced in the lower regions, while it does not yet play an important part at higher altitudes.

III. Maximum depths of snow cover

The minimum depths of snow cover are of special interest for winter sports, as these require a continuous duration of snow cover of a minimum depth, which depends on the conditions of the soil. Inasmuch as our records are sufficient for the treatment of small snow depths, the results will be discussed in § IV concerning the duration of snow cover.

On the other hand the maximum depths are important for transport and for the construction of avalanche protection works. Just for the last named purpose we ought to know the maximum depth and furthermore the maximum mass of snow which we must be prepared for. In Table 4 the indications about the maximum depth of snow cover of the 20 stations are summarized; the left part of the table contains the mean figures of the monthly maximum, calculated from the 30 monthly maxima of the observation period, the last column the absolute maximum which has been observed during the period at every station. This period of 30 years will be sufficient for a first orientation.

Table 4. — *Maximum depths of snow cover in cm (1913-1943) for December to March.*

Valleys and stations	Monthly average maximum				Absolute Max. 30 years
	Dec.	Jan.	Feb.	March	
<i>Prätigau</i>					
Landquart	23	36	28	12	115
Schiers	35	56	55	45	120
Küblis	43	64	66	51	135
Klosters	67	99	104	94	200
Cavadürli	77	110	120	120	210
Laret	92	132	145	151	275

<i>Davos Valley</i>					
Filisur	35	49	46	32	100
Glaris	59	87	89	85	180
Davos-Platz	60	88	97	90	195
Wolfgang	76	109	122	125	205
<i>Albula Valley</i>					
Bergün	43	62	64	57	125
Preda	60	77	92	96	170
<i>Engadine</i>					
Schuls	44	63	65	53	140
Zernez	44	64	69	61	180
Zuoz	53	70	79	77	165
Bevers	55	68	81	80	175
St. Moritz	58	75	86	89	185
<i>Fore Rhine Valley</i>					
Ilanz	37	55	55	39	138
Truns	50	79	78	59	167
Disentis	49	76	76	69	180

The results of the investigation of the maxima of the depth of snow cover may be summarized as follows :

1) The maxima of the depth of snow cover increase with altitude in every valley, the increase is somewhat parallel to the increase of the mean figures of the depth of snow cover. So the exterior valleys of the Grisons, which show the greater mean depths of snow cover, have also the greatest maxima, since in the inner valleys the maximum depths are much smaller.

2) For the climatological average there exists quite a narrow relation between the mean monthly maximum depths and the average figures of the depths of snow cover, as can be controlled by a comparison of Table 2 and 4 : the difference between the monthly maxima and the mean monthly depths of snow cover is exactly 20 cm for each month and the greater part of the Grisons ; only for the region from Klosters to Wolfgang with its abundant snow masses and for the Fore Rhine Valley this mean difference reaches almost 30 cm. From this rule it may be possible to estimate the mean maxima of snow depth for regions where the mean depth has been observed during a certain period.

3) For the calculation of avalanche protection works or other constructions the knowledge of *mean* maximum depths of snow cover will not be sufficient, but indications about the *absolute* maximum of snow depth to be expected will be necessary. These figures are given in the last column of Table 4 for a period of 30 years which will be sufficient for a practical evaluation of the snow masses which can occur. We find that in every valley the absolute maximum is, of course, increasing with altitude ; but this increase is different for the various valleys so that the maximum is smaller for the valleys with small snow depths. There exists quite a narrow correlation of the absolute maximum not with the altitude, but with the mean snow depth. If we design with A the mean depth of snow during the four months December to March (last column of Table 2) and with M the absolute maximum of snow depth to be expected at the same place, then

$$M = A + 70 \pm 20 \text{ cm.}$$

For any place of our region the absolute maximum of snow depth is 70 ± 20 cm greater than the mean depth of winter ; this formula is quite exact and is valid for the different valleys of Grisons with the same constant. It enables us to calculate from snow observations of a short period of some years only, the absolute maximum which can occur at the same place.

It is evident that for some technical problems the depth of snow cover must be multiplied by the probable specific weight of the snow, and this increases with the depth of the snow cover. So, for a double depth of snow cover in any comparison the total mass of snow will be increased nearly three times.

IV. The duration of snow cover

The duration of snow cover is a climatological element of great importance. Though the snow cover represents generally an obstacle to transport in mountainous regions, it can, however, sometimes facilitate traffic with isolated regions. On the other hand for winter sport resorts the snow cover represents an indispensable condition, whose quality and duration is of primary importance. The problem of the duration of snow cover has been discussed especially by Austrian scientists; a survey of these papers has been published by Steinhäuser (7).

Generally the duration of snow cover is defined by the number of days the soil is constantly covered with snow to a depth of at least one cm. For our records this limit cannot be applied, because the small snow depths were, as explained above, not observed exactly enough. Therefore we take into consideration here the duration of snow cover of at least 20 cm. For practical problems this limit, generally, is more interesting too, because a snow cover of 20 cm may already be troublesome to railway traffic and perhaps just sufficient for skiing of a kind.

The duration of a snow cover of such depth varies greatly for different winters; consequently it is necessary to adopt, for a climatological comparison of different stations, the same period for all stations. Therefore only 16 winters are taken into regard here.

The mean dates for beginning and end and for the duration of snow cover of at least 20 cm depth are given in Table 4 for the stations of Table 1.

Table 4. — Date of beginning and end and duration of a snow cover of ≥ 20 cm.

Valley and station	Altitude m	Beginning		End		Duration days	Mean date
<i>Prätigau</i>							
Landquart	523	Dec.	29	Jan.	29	31	Jan. 14
Schiers	657	Dec.	11	March	3	82	Jan. 21
Küblis	813	Dec.	11	March	11	90	Jan. 25
Klosters	1194	Nov.	29	April	2	124	Jan. 30
Cavadürli	1355	Nov.	17	April	17	151	Feb. 1
Laret	1525	Nov.	14	April	26	163	Feb. 4
Wolfgang	1629	Nov.	15	April	21	157	Feb. 2
<i>Davos Valley</i>							
Filisur	1084	Dec.	19	Feb.	27	70	Jan. 23
Glaris	1457	Dec.	8	April	6	119	Feb. 6
Davos-Platz	1543	Nov.	23	April	10	138	Jan. 31
Wolfgang	1629	Nov.	15	April	21	157	Feb. 2
<i>Albula Valley</i>							
Filisur	1084	Dec.	19	Feb.	27	70	Jan. 23
Bergün	1374	Dec.	14	March	26	102	Feb. 3
Preda	1792	Nov.	28	April	20	142	Feb. 8
<i>Engadine</i>							
Schuls	1290	Dec.	15	March	18	93	Jan. 31
Zernez	1474	Dec.	22	March	26	94	Feb. 7
Zuoz	1695	Dec.	4	April	9	126	Feb. 5
Bervers	1714	Dec.	1	April	11	131	Feb. 4
St. Moritz	1778	Nov.	30	April	10	131	Feb. 3
<i>Fore Rhine Valley</i>							
Landquart	523	Dec.	29	Jan.	29	31	Jan. 14
Ilanz	702	Dec.	18	Feb.	28	72	Jan. 23
Truns	855	Dec.	12	March	9	87	Jan. 25
Disentis	1133	Dec.	15	March	12	87	Jan. 28

(7) F. STEINHAUSER, Die Schneehöhen in den Ostalpen und die Bedeutung der winterlichen Temperatur inversion. Arch. Met. Geoph. Biokl. Ser. B 1, 63 (1948).

The duration of snow cover is dependent mainly on the following three factors :

- (a) altitude,
- (b) mean snow depth,
- (c) snow depth of the various winters.

The effect of the first two influences can easily be recognized with the curves of Fig. 4.

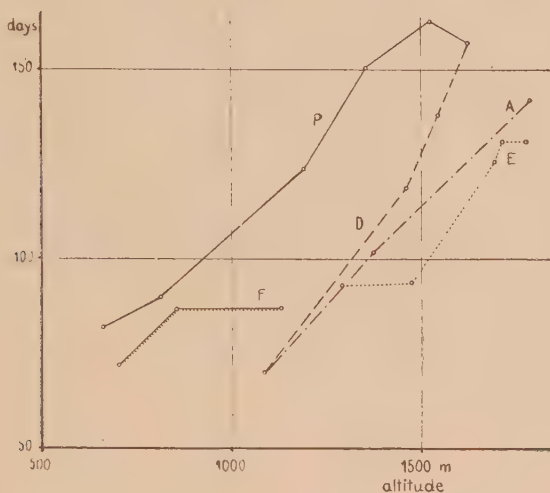


FIG. 4. Duration of a snow cover of ≥ 20 cm., as function of altitude.

In each of the valleys we can establish a clear increase of the duration of snow cover with altitude. In the Prätigau, the Davos and the Albula valley this increase is a linear function of altitude, since in the other two valleys the relation is more irregular on account of local influences. The snow cover duration at Landquart has not been used for the diagram, because it seems to be 30-40 days too low, doubtless as a consequence of a local influence of foehn.

It is evident that the duration of snow cover increases with altitude to a similar degree as the depth of snow cover. The deeper the depth of snow, the longer will be its duration. Just on account of this dependence on snow depth, the duration of snow cover is very unequal in the different valleys, even for the same altitude. The valleys with the smaller snow depths have also the shorter duration of snow cover. But the quotient between duration of a snow cover of ≥ 20 cm and e.g. the mean winter depth of it (according to Table 2) is not constant at all; it diminishes regularly from lower to higher altitudes and it is small for valleys with great snow quantities (as Prätigau) and higher for the valleys with relatively small snow masses (as Albula Valley and Engadine). Therefrom we can conclude that the duration of snow cover does not depend only on snow depth, but also on temperature which is closely related to altitude.

Another problem to be studied is the question whether the differences in the duration of snow cover are in the main the result of differences at the beginning or at the end of the period the snow cover lies. By an analysis of the data we find that with increasing altitude the snow cover generally begins to lie at an earlier date and lasts longer. The two effects are approximately symmetrical, at least for altitudes above 1000 m. The centre of symmetry (see Table 4) is 1st February; this date can then be regarded as the middle of winter. For lower altitudes the melting enters quite earlier, so that the centre of winter lies here one or two weeks earlier than in higher altitudes.

A good comparison between the different valleys can be obtained by means of Table 5, where the mean duration at an altitude of 1500 m and the increase of duration for an elevation of 100 m is indicated for the various valleys.

Table 5. — Mean duration of a snow cover of ≥ 20 cm at an altitude of 1500 m.

Valley	Duration days	Mean increase in days /100 m
Prätigau	160	9
Davos Valley	128	15
Albula Valley	116	10
Engadine	98	10

This table shows clearly that the interior valleys of the Alps with their smaller snow masses have much a shorter duration of snow cover than the exterior valleys; even for the same altitude the duration in the Engadine is more than two months shorter than in the Prätigau. The mean increase of duration is generally 10 days for an elevation of 100 m; only in the Davos Valley this gradient is much greater because the basis point of this valley (Filisur) is situated in a region with small snow masses, whilst the top approaches a region with deep snow cover.

As the mean duration of snow cover depends on the mean snow masses, it is evident that in individual winters the duration of snow cover is also greatly dependent on the quantity of snow; in winters with much snow the duration of snow cover naturally is greater than in winters poor in snow. But this increase of duration is not proportional to the depth of snow, it is an exponential function of the depth of snow cover and tends to a maximum value which cannot be exceeded for each altitude. This relation is represented by the figures of Table 6 and by Figure 5.

Table 6. — Mean duration of snow cover for different snow depths (in days).
(Valid for all valleys of Grisons)

Altitude in m	cm snow cover (winter mean)				
	20	40	60	80	100
500 - 800	60-70	100			
800 - 1200		105	125		
1200 - 1500	70-75	110	140	150-155	160-165
1500 - 1800		120	150	160-165	170-180

The duration of snow cover depends on two factors. The main influence is that of snow depth caused by the fact that a deeper snow cover needs a longer time for melting. Only a secondary effect is the increase of duration of snow cover with altitude; it is a consequence of the mean decrease of temperature with altitude, so that the snow cover has a greater resistance to melting and needs a greater heat of fusion.

For the figures in Table 6 no remarkable differences between the different valleys have been found, in spite of their very unequal depths of snow cover. A mean snow depth of 60 cm may for example be relatively low for one place and high for the other, the mean duration of this snow cover will then be dependent only on altitude. But, als clearly is shown by Fig. 5, the variability of duration of snow cover (at a given altitude) is much greater with the variations of snow depth than with increasing altitude (for a constant snow depth).

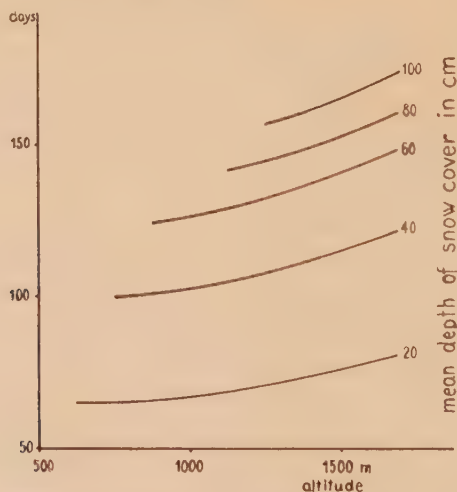


FIG. 5. Duration of snow cover in dependence on altitude and snow depth.

19) SUR L'ACTIVITÉ DE L'INSTITUT FÉDÉRAL POUR L'ÉTUDE DE LA NEIGE ET DES AVALANCHES AU WEISSFLUHJOCH SUR DAVOS, SUISSE

par

EDWIN BUCHER et ses Collaborateurs

En Suisse, comme dans d'autres pays d'ailleurs, c'est de la pratique que fut donnée l'impulsion d'une étude systématique de la neige.

Les dégâts causés par les avalanches et les ouvrages de protection contre ces dernières ont engendré de telles sommes que les modestes résultats de ces études ont déjà permis de réaliser d'importantes économies.

La commission suisse pour l'étude de la neige et des avalanches, fondée en 1931, se concentra tout d'abord sur le problème pratique de la protection contre les avalanches. De courtes périodes d'observation suffirent à démontrer la nécessité de considérer le problème dans son ensemble. Après quelques investigations à Davos, un laboratoire primitif fut établi au Weissfluhjoch à 2660 m d'altitude. Il fut remplacé en 1942 par un institut spacieux. Après la guerre c'est une joie bien compréhensible pour les chercheurs de l'institut que de pouvoir communiquer à d'autres nations les résultats des recherches effectuées en Suisse sur la neige et sur les avalanches.

L'activité de l'institut se divise en deux groupes : 1) les recherches fondamentales et 2) leurs applications.

A. Recherches fondamentales

1. *La météorologie et l'hydrologie* sont les tâches primordiales des recherches générales sur la neige. Le genre, l'épaisseur et les changements de la stratification de chaque couverture de neige sont régies uniquement par les conditions météorologiques. En voulant entreprendre la mesure des précipitations atmosphériques, du vent, de la température, de la radiation, il fallut se rendre à l'évidence que les appareils utilisés ordinairement étaient impropres à l'usage en montagne. La principale difficulté consiste à mesurer les précipitations sous forme solide (neige, etc.) Une différence importante existe entre les données des pluviographes et des totalisateurs et la valeur d'eau effective de la couverture de neige. D'après les recherches effectuées jusqu'à maintenant il semble que la notion ordinairement admise, de l'augmentation des précipitations avec l'altitude au cours des mois d'hiver par opposition aux périodes estivales, soit mal fondée.

Les facteurs importants influençant l'évolution des couches sont : l'ampleur des précipitations, la force du vent et la température. D'importantes chutes de neige par température élevée s'accumulent généralement en strates compactes. Le vent a la même influence, il tasse les couches. De petites chutes de neige à basse température se métamorphosent intensivement en gros grains et la cohésion entre les cristaux tend à s'amoinrir.

2. *Investigations générales sur le terrain*

Les observations de l'évolution des couches de neige relevées chaque année sur le champ d'essai standardisé du Weissfluhjoch ont une triple importance :

En premier lieu le prélèvement périodique de profils de la résistance au battage, de température et de la coupe stratigraphique, indique la relation existant entre le temps et la métamorphose de la neige. Ces observations constituent les bases de l'étude systématique des conditions de neige dans toutes les Alpes. De plus la continuité de ces prélèvements permet d'établir des données statistiques intéressantes sur le climat ; comme par exemple sur la moyenne et l'amplitude de la durée d'enneigement, sur l'épaisseur de la couche de neige, sur la quantité de la neige fraîche, etc.

Depuis 1936 ces prélèvements furent effectués systématiquement au même endroit et permettent d'établir les valeurs moyennes de l'enneigement, à 2540 m.

d'altitude, de dix années consécutives, soit des hivers 1936-37 à 1945-46. La durée moyenne d'enneigement est de 265 jours, soit huit mois et demi et s'étend du milieu d'octobre à la première décade de juillet.

D'après les moyennes établies mille mètres plus bas, à Davos par le Dr. Mörikofer, la durée de l'enneigement n'est que de 172 jours soit 5,7 mois.

La courbe moyenne de la hauteur de la neige du champ d'essai du Weissfluhjoch monte régulièrement jusqu'au maximum de 220 cm le 10 avril puis elle redescend légèrement en s'incurvant vers le bas jusqu'à la fonte complète de la première décade de juillet. La plus grande épaisseur de neige observée pendant cette période est de 365 cm, mesurée le 8 mars 1946.

En complément des profils du champ d'essai, des échantillons de neige de chaque couche sont analysés au laboratoire dans le but de déterminer la métamorphose des grains en fonction du temps, de même que le poids spécifique, la résistance à la rupture et la perméabilité à l'air des différentes couches.

Le tassement de chaque couche est mesuré électriquement par un dispositif utilisant le principe du pont de Wheatstone.

Pour l'étude du problème de l'état d'humidité ou de la contenance d'eau de la couverture de neige, des mesures d'écoulement de l'eau de fonte d'une surface donnée sont effectuées chaque année en plus de la détermination périodique de la valeur d'eau de cette couverture de neige. Afin d'établir l'influence de l'altitude, des prélèvements identiques sont effectués à Büschalp 1960 m, à Davos 1545 m, à Laret 1530 m, et à Klosters 1200 m.

Pour déterminer l'influence de la durée d'insolation et de son intensité sur l'évolution de la couverture de neige et sur la métamorphose des cristaux constituant la neige, deux champs d'essais furent observés à Davos au cours de l'hiver; un au soleil, et l'autre complètement à l'ombre. Simultanément l'influence des forêts sur la couverture de neige est étudiée à Davos et à Laret. Toutes ces observations tendent à expliquer les variations de l'évolution de la couverture de neige.

L'étude du rampeement de la neige sur le sol à été entreprise en divers endroits, entre 2660 et 2000 m d'altitude, à ciel ouvert et sous bois. Les méthodes de mesure employées sont encore trop approximatives pour que l'on puisse déjà tirer des conclusions générales.

2. *Recherches physiques et cristallographiques.*

Cette section comporte en premier lieu l'étude de la métamorphose. Les cristaux changent de forme avec le temps; les étoiles primitives deviennent des grains plus ou moins gros. Le mélange des grains, analysé auparavant dans une série de tamis, se fait maintenant dans un appareil où les particules de glace tombent dans un liquide. Les plus gros grains arrivent les premiers sur une balance retenue par un ressort dont le mouvement est fixé photographiquement. Cet appareil enregistre la courbe caractéristique du mélange des grains d'un échantillon et cette méthode serait certainement applicable pour d'autres matériaux que la neige.

La métamorphose des étoiles de neige fraîche a été étudiée d'après des séries de photographies sur lesquelles on remarque distinctement la disparition progressive de la forme dentritique primitive. Cette métamorphose peut être fixée par un enregistrement cinématographique au ralenti. Le film pris au printemps 1947 montre quelques phases intéressantes de la métamorphose du cristal.

La détermination de la contenance de chaleur de la couverture a une grande importance aussi bien pour la mécanique de la neige, que pour le développement saisonnier des couches et également pour l'hydrologie et la biologie des plantes sous conditions hivernales. Après les mesures effectuées au cours de l'hiver 1945-46, sur les variations de température dans la couverture de neige, la détermination de la conductibilité de la neige à la chaleur fut entreprise. On trouve dans la littérature la relation entre le coefficient de conductibilité de la chaleur et le poids spécifique. Pourtant ces valeurs varient et ne tiennent pas compte de la grosseur des grains. La part de chaleur transportée par le chassis de glace constituant la neige d'une part et par les pores d'autre part ne se trouve pas dans la littérature du sujet. Des mesures spéciales furent donc nécessaires.

La contenance de chaleur de la couverture de neige est en relation étroite avec l'échange de chaleur entre la neige et l'air. Des observations antérieures ont montré que l'évaporation et la condensation se font fort différemment à l'ombre ou au soleil.

D'après ces expériences, un dispositif de mesure au soleil fut installé au Weissfluhjoch. Au cours des quatre premiers mois de l'année 1947, il fut possible de faire des mesures successives d'évaporation à l'ombre et au soleil. Les facteurs météo-

rologiques (température vitesse du vent, humidité, radiation, etc.) sont enregistrés simultanément, de sorte qu'il est possible d'établir le bilan de l'évaporation pour la région du Weissfluhjoch. Si des mesures irréprochables des précipitations et de l'eau de fonte pouvaient être obtenues, on aurait une vue d'ensemble de la contenance d'eau sous forme de neige de la couche de neige d'un hiver.

Le freinage de l'écoulement de l'eau à travers la couverture de neige, peut être défini par le temps séparant la précipitation sous forme de pluie sur une couche de neige d'une épaisseur donnée et le début de l'écoulement à la base de cette couche. Les unités suivantes ont été choisies; amenée d'eau de 10 millimètres de hauteur par minute, couche de neige d'un mètre d'épaisseur. La retenue varie suivant le genre de neige de 3 à 7 minutes. L'écoulement se continue pendant un certain temps après l'arrêt de la pluie. Cet écoulement à retardement diminue de plus en plus et se poursuit toujours plus longtemps que la durée du freinage. Les essais se font en arrosant d'eau glacée un cylindre de neige. La quantité d'eau tombée et l'écoulement à la surface inférieure s'enregistrent automatiquement sur un tambour. A l'aide de ces mesures, il est possible d'évaluer le freinage de l'écoulement de la pluie sur un pente couverte de neige ou même sur toute une région si l'on connaît les caractéristiques de la couverture de neige.

Dans un domaine plus mécanique les déformations plastiques et élastiques de la neige sous l'influence de forces extérieures furent déterminées, au moyen d'échantillons soumis à la torsion et dont les très petits mouvements sont mesurés par un dispositif à miroir.

3. *Recherches mécaniques.*

Les principales investigations mécaniques consistèrent à déterminer la viscosité et la résistance à la rupture des différents genres de neige. Ces deux propriétés varient principalement avec le poids spécifique, la grosseur moyenne des grains et la température, et aussi avec les surfaces de contact d'un grain à l'autre.

En principe la neige se comporte comme les métaux au voisinage de leur température de fusion, qui furent l'objet de nombreuses études. D'emblée ce fait paraît étonnant, mais il est immédiatement compréhensible si l'on analyse les caractéristiques communes des deux matériaux: la structure cristalline et la température voisine du point de fusion. Les expériences sur la résistance et la plasticité se basent donc sur les méthodes utilisées dans la métallurgie. Les caractéristiques de la neige dépendent aussi de sa porosité. Le résultat le plus important est l'explication de la rupture. Les endroits les plus faibles d'un échantillon de neige se trouvent aux points de contact d'un cristal à l'autre et le matériau se brise à la limite des grains. Ce fait nous introduit dans la relation existante entre la structure et la résistance de la neige et on peut en déduire certaines notions quant à la variation de la cohésion en fonction de la forme du grain, du poids spécifique de la neige et de la température. La neige constitue un aggrégat dont les grains sont entourés d'une couche visqueuse, qui peut transmettre les efforts sur leurs voisins. Représentons nous les énormes variations possibles de ces grains quant à leur grandeur, leur forme et les particularités de leur surface; si la solidité des contacts d'un cristal à l'autre varie sous des influences cristallographiques, en fonction de la température, la diversité de la résistance de la neige devient compréhensible.

La plasticité du matériau s'explique de même manière: sous l'influence de forces extérieures, la déformation de la neige se fait en premier lieu par déplacements à la limite des grains et d'une façon très restreinte par translation interne dans les cristaux. Les facteurs principaux influençant la plasticité sont: la forme du grain, le poids spécifique de la neige et la température. Les essais donnent des valeurs du coefficient de viscosité η variant de 10^6 à 10^{11} kg sec m^{-2} pour de la neige fraîche soit de 10^8 à 10^{13} poises.

Les courbes de freinage de la vitesse de rampement de la neige sur un plan incliné et butant contre un mur furent déterminées au laboratoire.

La question primordiale qui se pose est de savoir jusqu'où s'étend la zone de retenue. Les essais ne donnèrent pas de résultats probants. Les mouvements des points mesurés deviennent rapidement trop petits pour que le résultat puisse être suffisamment précis. Si l'expérience se poursuit pendant plus longtemps, la non-homogénéité du matériau fausse l'image normale de l'écoulement. Ces essais doivent être continués avec des méthodes de mesures plus précises. Les écoulements se faisant dans des conditions limites égales sont plus simples à observer. Des blocs de neige de $50 \times 50 \times 20$ cm furent placés dans un cadre de façon à ce qu'une des diagonales de la grande surface soit verticale. Le chemin parcouru à la suite du tassement et du cisaillement d'une série de repères répartis sur la surface est mesuré, puis

comparé avec les hypothèses théoriques. Une étude plus poussée permettra probablement de résoudre le problème du calcul de l'écoulement et des tensions intervenant dans la neige ou la glace dans des conditions limites compliquées.

Le praticien s'intéresse surtout aux poussées que doivent supporter les constructions. Ce sont les efforts agissant sur la tranche fixe de la couverture de neige. La pression de la neige ne fut mesurée auparavant que dans la nature (Weissfluhjoch Alp Grüm), mais actuellement elles est étudiée au laboratoire. Une couche de neige repose sur une planche inclinée à 45° et doublée d'un treillis annulant le rampe-ment au sol.

Trois lamelles d'anticorodal disposées horizontalement les unes sur les autres et retenues par des ressorts permettent, non seulement de mesurer la poussée de la neige, mais encore la répartition de cette poussée suivant la hauteur. Quelques problèmes de tassement et de frottement furent aussi étudiés. La pression de tassement c'est à dire une rapide compression de la neige rentre dans cette catégorie. Si le poids spécifique de la neige augmente, des forces toujours plus grandes sont nécessaires pour comprimer le matériau. Cette fonction est représentée par une courbe exponentielle c'est-à-dire que les forces tendent vers l'infini pour l'agrégat à porosité minima. La température et la grosseur des grains ont la même influence que pour les autres essais de résistance. L'intégration des forces entrant en action donne la résistance totale à vaincre pour comprimer la neige.

Dans un mouvement glissant (skis, traineaux), le frottement de glissement s'ajoute au frottement de tassement, et dépend de la qualité du patin, de la charge spécifique, de la température, du genre de neige et de la vitesse du mouvement. Les résultats obtenus trouvèrent leur application pratique dans l'étude du déblaiement de la neige et du tassement des routes etc. exécutée pour le compte de la commune de Davos.

B. Applications

1. *Service des avalanches*

A la fin de la guerre, l'institut eut pour mission d'adapter le service des avalanches militaires aux conditions civiles. La tâche consiste à récolter de différentes parties des Alpes, des renseignements sur les conditions de neige et d'avalanches, permettant d'une part d'évaluer le danger éventuel et d'autre part d'obtenir ainsi une documentation scientifique. L'élaboration du bulletin d'avalanche se fait d'après les observations de 20 stations réparties dans les Alpes. D'autres postes encore ont un programme réduit et leurs données permettent l'étude de l'influence de l'altitude sur l'enneigement. Ce service des avalanches civil existe depuis 1945. Les bulletins diffusés par la presse et la radio ont trouvé un excellent accueil aussi bien parmi les hôtes étrangers qu'au sein de la population helvétique. De semblables efforts sont tentés en France, en Autriche et en Tchécoslovaquie.

2. *Constructions*

Les tensions existant dans la couverture de neige dépendent de la configuration du terrain, de la répartition de l'épaisseur de la neige et de la qualité de cette neige. Le but des constructions est d'influencer la répartition des tensions de façon à éviter une rupture. Pour cela, deux possibilités fondamentales existent : soit influencer le dépôt de la neige (constructions pour ou contre le transport de neige par le vent), ou bien retenir la couche de neige (constructions d'appui). La dissertation de M. Edwin Bucher : *Beitrag zu den theoretischen Grundlagen des Lawinenverbaues* (contribution aux bases théoriques des constructions contre avalanches) développe ce problème.

3. *Tassement et déblaiement de la neige.*

Sur la demande du syndicat d'initiative de Davos, le problème du trafic hivernal fut étudié. Pour le tassement de la neige sur les routes, trois rouleaux de différentes dimensions furent prévus dont chacun tasse successivement sa part. Le choix du rayon et du poids détermine la force de traction nécessaire. Après la passage du troisième rouleau, la neige est suffisamment compacte pour permettre le trafic automobile.

Les chasse neiges de petite dimension manquent aujourd'hui pour le déblaiement rapide des chemins et places de garage. Le prototype étudié d'une fraiseuse ventilateur à deux roues comble cette lacune et rendra certainement de bons services dans tous les centres alpins.

(Traduit de l'Allemand par André Roch)

INFLUENCE DE LA COUVERTURE NIVALE SUR LE RUISSELLEMENT ET SPECIALEMENT SUR LES INONDATIONS

par

M. PARDE

A. — Rôle excessif fréquemment attribué à la fonte des neiges

Ce problème mérite la plus vive attention pour trois motifs :

1) Il est bien vrai que certaines inondations ont pour cause unique, principale, ou non négligeable, des fontes de neige.

2) Une quantité de gens, non seulement dans le public non averti, mais parmi les personnes qui devraient mieux connaître les rivières, attribuent un rôle exagéré à la fonte des neiges, en tant que génératrice de grandes crues, et se contentent, à son sujet, de preuves très insuffisantes. Il leur suffit qu'une couche de neige se soit liquéfiée ou qu'on ait supposé, à cause de la date, son existence, puis sa fusion, pour que maintes personnes dénoncent ce phénomène vrai ou supposé comme facteur responsable.

3) Cette opinion devenue presque obligatoire, sorte de tarte à la crème lors des crues, est aussi tenace que celle qui attribue tous les méfaits des rivières, même de plaine, au « déboisement », même quand celui-ci n'a pas eu lieu sauf dans l'imagination des auteurs, journalistes ou autres, idée si automatique et si autonome, que l'on croirait à entendre certains de ces promoteurs, que le déboisement, diabolus ex machina, suffit à déclencher des malheurs hydrologiques, même quand il n'a point plu, ou que la neige n'a point fondu.

Pour en revenir aux fusions nivales, les explications des crues par elles résistent à toutes les tentatives d'analyse exacte ou approchée, et en particulier aux démonstrations arithmétiques les plus simples et les plus évidentes (élément de compréhension bien trop négligé),

Essayez, par exemple, de démontrer à un fanatique des fontes une vérité comme celle qui suit : soit une crue qui a mis en jeu de par son volume liquide total et la surface réceptrice, l'équivalent de 120 à 140 mm d'eau sur ce bassin. Vous avez vu fondre, ou vous savez qu'il a fondu 20 centimètres de neige récente, en moyenne sur la surface réceptrice. Cet apport a représenté au plus 25 à 30 millimètres d'eau, donc au maximum, de 18 à 25 % du volume de la crue. Dans un grand nombre de cas notre interlocuteur, même cultivé, hochera doucement la tête, et finira par rompre la discussion en vous déclarant, et en vous laissant entendre, que vous êtes un négateur d'évidence, puisque cette neige, il le sait, a bel et bien été liquéfiée. Et l'on n'a plus qu'à lever les yeux ou les bras au ciel et à fuir soi-même l'entretien en souhaitant que de plus en plus les gens épris d'hydrologie lisent, ou soumettent à leurs investigations personnelles, le plus possible de cas où le volume des crues, les chutes d'eau et les tranches nivales préexistantes auront été déterminés avec une approximation suffisante.

Ce que nous développons là n'est pas une boutade, mais correspond à nos souvenirs de maintes discussions et de méprises parfois décourageantes. Nous avons lu que la fonte des neiges sur le Canigou avait lourdement aggravé la crue célèbre des rivières du Roussillon en octobre 1940. Or, un simple examen de la carte et une méditation brève sur les premières informations et sur les données du problème auraient montré qu'à coup sûr il ne subsistait plus de neige du côté du Tech qui roula les débits les plus dévastateurs lorsque commencèrent les grandes averses. Sur le versant de la Tet, une couche s'étendait encore, mais représentait au plus 50 millimètres d'eau de fusion et probablement moins de 25, au-dessus de 2000 mètres, soit sur une petite partie des bassins les plus sinistrés qui reçurent en un seul jour plusieurs centaines de millimètres de pluie.

La-dessus notre contradicteur ne s'est point longuement obstiné. Mais nous avons trouvé plus d'entêtement dans le même sens après le désastre de septembre 1920, en Haute Maurienne. Entre deux averses tièdes, d'origine méditerranéenne, amenées sur la chaîne-frontière par la Lombarde, ou vent d'Est-Sud-Est, des incursions passagères d'air froid amenèrent au-dessus de 2000 mètres des chutes de

neige. Un technicien d'ordinaire mieux inspiré voulait que les laves foudroyantes des torrents l'Envers, l'Ambin, le Charmaix, eussent été déchaînées en grande partie par la fonte de cette neige.

Le raisonnement, des connaissances climatiques sur les phénomènes nivaux de septembre, et la considération grossièrement approximative des débits eussent montré suffisamment que, selon toutes chances, les laves avaient eu pour cause essentielle des averses diluviennes dans les bassins supérieurs, avec un très faible apport de fonte. J'essayai de développer cette explication d'autant plus vraisemblable que je venais de recevoir de mes correspondants italiens des chiffres suggestifs sur les précipitations reçues à la même date par le versant piémontais des Alpes occidentales : à savoir, jusqu'à 200 et 300 millimètres en 1 jour à quelques stations qui ne furent sans doute pas les lieux les plus arrosés en la circonstance. Le contradicteur, qui savait par des témoins et des journaux, qu'il y avait eu de la neige, qui avait son explication et qui n'en demandait pas plus, refusa de se laisser ébranler. Il finit par me dire que j'étais un de ces théoriciens qui jugent des phénomènes naturels seulement d'après les livres. Evidemment ! Et celui qui voudrait discourir des averses et des crues, d'après son expérience personnelle et directe, aurait une faible base documentaire. Car les opinions sur ces choses ne commencent à devenir fondées que lorsqu'on peut expliquer au minimum, quelques dizaines de grandes crues. Et celui qui s'est trouvé, au maximum, plus de 2 ou 3 fois sous l'averse et dans les gâchis de fusion lors de ces cataclysmes, peut se considérer déjà comme très favorisé (si l'on peut dire ainsi) par la chance, et pour énoncer carrément les choses, il serait bon que ceux qui raisonnent et tranchent ou rient sur les crues aient une certaine érudition livresque (on s'excuse d'employer ces mots décriés) sur ces phénomènes et sur leurs causes. Ceci revient encore à déclarer, en termes propres, que la plupart des témoins en la matière ne peuvent guère voir plus loin que le bout de leur nez ; et s'ils ont pu et su étendre ce champ d'observation directe à quelques kilomètres, on conviendra qu'il est encore bien restreint. Montrons, par un nouvel exemple, comment sa faible amplitude peut égarer celui qui en réalité n'a rien vu sur place, outre des phénomènes locaux parfois insignifiants.

Lors des crues dites océaniques du Rhône, avec pluies amenées par les vents de Sud-Ouest à Ouest, Lyon, en 3 ou 4 jours grâce à l'abri du Massif Central, ne reçoit que des abatements d'eau infimes : par exemple, 3 à 10 millimètres. Et même en admettant un certain accroissement des chutes vers le Jura et les Vosges, un non initié à ce type de temps, ne peut savoir qu'aux mêmes dates, il tombe facilement 120 à 200, voire 250 millimètres à St-Claude, à Morez, aux Rousses, Samoëns, à Thonon, à St-Pierre de Chartreuse, et que c'est de ces déluges que résulte la crue non de la fonte de 10 centimètres de neige dans l'avant-pays de 20 à 40 centimètres au-dessus de 500 ou de 1000 mètres sur les montagnes : apport parfois plus ou moins compensé par la rétention de fortes quantités pluviales sous forme de neige, au-dessus de certaines limites d'altitude, comme on le verra ci-dessous.

Parfois encore, les témoins raidis par leur préjugé (que l'on pourrait bien qualifier alors de préconçu... à l'avance) se refusent à reconnaître que la pluie est tombée sur eux-mêmes, et à en interpréter les conséquences. Ainsi, lors de la crue lorraine désastreuse du 25 au 31 décembre 1947, divers observateurs ou commentateurs, apprenant que l'avant-pays lorrain désenneigé avait reçu après 30 à 50 millimètres de chutes préparatoires, 65 à 80 millimètres en un jour, ne paraissent pas avoir songé un instant que cette pluie pouvait avoir été la grande responsable de la catastrophe. Il leur a fallu invoquer comme grande coupable la fonte des neiges sur les Vosges. Effectivement ce facteur a joué, et pas de façon mesquine.

Il paraît avoir liquéfié jusqu'à 50 ou 60 centimètres vers 1000 mètres, peut être 30 en moyenne au-delà de 500 ou de 600 mètres, soit une tranche d'eau moyenne de 30 à 70 mm au plus, à ce qu'il semble, mais sur quelques 900 km², alors que la surface réceptrice totale au confluent de la Meurthe et de la Moselle dépasse 6500 km² ; et que la zone intéressée par la fusion a reçu jusqu'à 180 millimètres de pluie en un jour, à certains postes (1) et plus en quelques jours consécutifs. Même au pied des Vosges, sur la Moselle à Epinal par exemple, il est clair, d'après ces chiffres, et sans qu'on se livre à une analyse arithmétique plus poussée, que la neige a joué en l'occurrence un rôle de second plan. A Nancy, Toul et plus nettement encore à Metz, son action a été plusieurs fois, peut être 10 à 12 fois moindre que celle de la pluie.

(1) Ceci d'après un mémoire excellent de M. ROTHÉ, Directeur de l'Institut de physique du Globe de Strasbourg, et d'après des renseignements que nous a fournis la Météorologie Nationale.

B. — Etudes nécessaires sur le rôle de la neige dans les crues

Après avoir lu ces remarques, dont trop d'entêtements expliquent la vivacité, on nous demandera comment faire pour évaluer avec une suffisante approximation, le rôle de la neige dans les crues. Notre réponse de principe sera brève, d'ailleurs beaucoup plus facile que la mise à exécution du conseil : Il faut tâcher de déterminer les volumes ruisselants d'origine nivale d'après des observations et par les voies de l'arithmétique.

Il faudrait d'abord qu'à toutes les stations pluviométriques on n'oublie jamais de marquer par une astérisque les chutes d'eau tombées sous forme de neige. Puis les observateurs devraient avoir la mission expresse et bien contrôlée de marquer chaque jour les épaisseurs de la couche nivale totale sur le sol. Aux jours où le vent fléau de la nivométrie, aurait créé des inégalités dans l'épaisseur du manteau, l'observateur devrait inscrire un chiffre qu'il jugerait conforme à la moyenne non perturbée par le vent, pour l'ensemble de son petit secteur. Il nous semble que tout homme doué de bon sens et d'esprit d'observation, et bien conseillé à l'avance, devrait être capable de donner là-dessus, une approximation suffisante.

Puis, dans les régions où l'on aurait lieu de craindre des crues sérieuses dues à la fonte des neiges, il serait bon que de temps en temps, et notamment lorsqu'un redoux pluvieux se dessinerait, on mesurât l'épaisseur de la tranche de neige, et son équivalent en eau de fusion à une quantité d'endroits, aux stations pluviométriques ordinaires, et ailleurs, en des lieux typiques, déterminés à l'avance, toujours les mêmes. On pourrait prélever la neige au moyen de tubes sondeurs, en métal léger, comme le fait l'*Electricité de France* pour divers bassins montagneux, sur l'exemple de l'illustre professeur américain, M.J.E. Church et de ses collaborateurs.

En outre, après chaque grosse crue accompagnée de fonte on devrait faire des enquêtes sur place pour déterminer, d'après les souvenirs des habitants, quelles quantités nivales étaient disponibles avant les phénomènes.

Nous verrons bientôt que dans les bassins très accidentés de haute altitude, les réserves de neige, quoique bien plus abondantes qu'en plaine ou dans les petites montagnes donnent lieu d'ordinaire à de gros débits continus, mais rarement à des inondations graves. Les zones où des combinaisons très redoutables de pluies d'averses et de neige fondue peuvent se produire sont en somme assez restreintes. Parmi elles, nous citerons les Appalaches du Nord, et la crue mémorable de mars 1936 qui y a sévi. Après ce cataclysme, les ingénieurs du *Geological Survey*, grâce à des enquêtes approfondies ont pu assez bien retrouver les équivalents en eau de la neige préexistante en une foule de points. Puis ils calculèrent les moyennes pluviales lors des deux averses successives qui eurent lieu, à une semaine d'intervalle (la seconde étant de beaucoup la plus grave) et les volumes liquides totaux engendrés par chacune des averses. Sur ces bases et sachant à peu près quels quotients d'écoulement il fallait adopter pour les seuls débits pluviaux, ils ont pu déterminer assez bien les parts respectives de la neige fondue dans chaque intumescence et sur les divers bassins (1).

Par la suite d'autres chercheurs (2) du même service ont refait les calculs pour un petit nombre de ces rivières (Pennsylvanie et Etat de New York). Ils ont recalculé les moyennes pluviales et tenu compte de la neige non fondue subsistant sur le sol à la fin des intempéries, élément que bien entendu il convient de ne jamais oublier.

C'est seulement après des investigations et supputations de ce genre que l'on peut émettre une opinion point trop aventurée sur la part de la neige dans le ruissellement. D'après ces études et quelques légères rectifications que nous avons cru devoir tenter, on peut présenter sans trop d'inexactitude les conclusions suivantes :

Pour le Merrinack à Lowell, et pour le Connecticut à Hartford durant la phase principale, la neige a pu fournir respectivement 45 à 50, et 55 à 60 % des volumes totaux. Plus au Sud, contrairement à l'opinion populaire initiale, le rôle de la fonte a été bien moindre ; par exemple, pour la Susquehanna un sixième au plus, pour le Potomac presque rien. Notons que les débits maxima spécifiques à l'issue de bassins comparables furent plus considérables sur le Susquehanna et sur le Potomac que sur les rivières de la Nouvelle Angleterre. C'est que les réseaux des premières sont plus inclinés et plus concentrés. Mais les maxima des cours d'eau du Sud ne dépassèrent pas de façon excessive les records précédents tandis que le Merrimack l'emporta

(1) Water Supply Papers du Geological Survey, U.S.-Department of the Interior, Washington, N° 798, 799 et 800 (466, 380, 351 pages avec une abondante illustration).

(2) W.S. Paper du Geological, Survey, N° 915, 1947, 137 pages.

de 60 % par son débit maximum sur toutes ses crues antérieures et le Connecticut de 48,5 %. Ecart sensationnel et qui expliquent ces énormes ravages.

Des évaluations non dépourvues de signification peuvent encore être faites même sans connaissance directe du manteau nival existant avant la crue, si l'on obtient de bonnes connaissances sur la pluie qui a participé au phénomène, sur le volume total de l'intumescence et sur le coefficient d'écoulement rendu probable par la nature du bassin et par la date; tout l'excédent de débit non justifiable par la seule pluie provient de la fusion nivale. Plus est gros le cube liquide provenant de cette origine, par rapport à celui de ruissellement pluvial, moins grosse est l'erreur que l'on risque sur le rôle de la neige. Encore faut-il ne pas se tromper de façon grossière sur le coefficient d'écoulement. Or, en ce genre d'estimation, on peut se fourvoyer plus lourdement que je le supposais naguère. Par exemple, d'après des exemples très nombreux j'aurais estimé à 75 - 90 %, pour des bassins montagneux cristallins, le coefficient d'écoulement d'une averse exceptionnelle hivernale. Or, dans les montagnes ainsi constituées de Californie du Sud, en mars 1938, les rivières n'ont écoulé que 45 à 75 % d'averses atteignant les totaux monstrueux de 725 mm en quelques jours, 10 à 35 % en général de pluies de 175 à 300 millimètres qui, dans nos montagnes européennes ou dans les Appalaches, n'auraient subi que des déchets de 10 à 30 % au plus et occasionné des désastres. Donc dans les mêmes zones, les coefficients n'auraient pu atteindre 75 à 90 % pour 150 à 250 mm écoulés, que de manière factice, grâce à un apport de neige fondue vraisemblablement peu différent de celui des pluies.

C. — Faits généraux à retenir sur le rôle de la neige dans les crues

Les éléments de calculs ou d'évaluations de ce genre manquent bien souvent. Aussi avons-nous dû recourir beaucoup au simple raisonnement, pour aboutir sur les divers taux du renfort nival en crues aux conclusions provisoires suivantes :

1.—Crues dues à la fonte seule sur les rivières nivales et glaciaires de montagne

Elles sont régulières, chaque année, sur une foule de cours d'eau, mais rarement graves. Cependant, au début de la période principale de fusion les rivières de haute montagne : Isère, Arc, Arve, Rhône, Aar, etc. portent sur leur bassin dans leurs réserves nivales plusieurs centaines de millimètres d'eau de fonte : à savoir, de quoi engendrer des cataclysmes monstrueux si la liquéfaction ne faisait en quelques jours seulement. Mais c'est précisément ce qui est impossible parce que la différenciation des altitudes, donc des températures, échelonne sur des semaines et plutôt sur des mois, la transformation en eau ruisselante. Quand la neige fond en masse de 1500 à 2000 mètres, elle reste presque intacte de 2500 à 3000 mètres et ainsi de suite. On n'observe donc en général que de hautes eaux continues non dangereuses venant de zones de plus en plus élevées. Les rares crues nivales dommageables surviennent seulement quand la fonte ayant été tardive et le manteau étant resté très étendu jusqu'à des cotes assez basses, une chaleur intense se manifeste brusquement et attaque la neige et la glace sur une grande différence d'altitude donc sur une vaste superficie. On a noté dans ces conditions sur le Rhône alpestre, avant le Léman, des crues voisines des records (juillet 1897 et juin-juillet 1935). Mais il convient de rappeler qu'en ce bassin très abrité contre les grandes averses, les maxima connus sont en réalité peu imposants (200 litres-seconde au plus par km², pour 5.220 km²). Pour l'Isère à Grenoble, qui peut débiter 2000 m³ et plus, soit quelque 400-litres seconde, par km², la fonte ne semble guère apte à fournir plus des deux cinquièmes de ces valeurs, ce qui l'empêche lorsqu'elle seule grossit la rivière (juin 1919, 1937) d'occasionner une inondation grave. Seuls de petits bassins avec dénivellation pas énorme, peuvent parfois connaître une attaque soudaine de la chaleur sur une surface totale encore entièrement enneigée, et ainsi des gonflements sérieux, voire nuisibles. Ce fut le cas pour la Valloirette en Savoie, en Juin 1935.

2. — Crues à la fois pluviales et nivales

a. — Bassins de haute montagne

Les événements peuvent devenir bien plus graves en pleine période de fonte si la rivière étant déjà très grossie par la fonte des neiges, il survient une forte pluie. Le refroidissement engendré d'habitude par la chute d'eau en cette saison peut n'avoir réduit l'écoulement nival que dans une médiocre mesure lorsque survient la maximum de ruissellement pluvial. Alors on observe une crue combinée redoutable.

Mais ceci, bien que réalisable logiquement n'est souvent que théorie. Nous ne connaissons pas d'exemple d'une très grande crue de l'Isère due à ce mécanisme. Et dans les gonflements que le Rhône alpestre a dus à cette cause mixte (juillet 1883 et 1902); aucun n'a été réellement formidable (par exemple 1500 à 2000 m³, contre des crues pluviales ou nivales pures de 800 à 1000 m³).

Par contre, selon toute vraisemblance, il faut attribuer à la neige non pas la responsabilité unique, mais un rôle très notable dans la genèse de plusieurs des déluges qui ont inondé Grenoble par le fait de l'Isère (novembre 1651, 1859, décembre 1740, peut être octobre 1778). Cependant, faute de données précises sur ces phénomènes anciens, nous ignorons quelles proportions respectives d'eaux nivales (25, 40, 50 % ou plus) y ont participé. En outre ces combinaisons pluvio-nivales ne paraissent point possibles, du moins avec une telle gravité, dans tous les bassins alpestres. Il faut une situation, une latitude et une distribution altimétrique, telles que dès le début de l'automne, un vaste apport de neiges précoces puisse s'étendre très bas, et que peu après, avant le cœur de l'hiver, des averses tièdes se déchainent et s'accompagnent de fonte sur une vaste superficie relative; par exemple, deux tiers ou plus du bassin. Ce fut plusieurs fois le cas précisément pour l'Isère. D'ailleurs aucune de ces concomitances graves, n'a été notée depuis 1859, mais pour un rien le drame se fût reproduit en 1939. Si la couche de neige dépassée sur le bassin, jusque dans le Gresivaudan et la Combe de Savoie (74 mm d'eau de fusion à St-Pierre d'Albigny) avait attendu pour fondre, ce qui n'était pas impossible, la pluie tiède du 16 au 17 novembre suivant, on aurait assisté sans doute à un désastre.

b. — Bassins de petites et moyennes montagnes, ou de plaines en Europe occidentale

D'autre part, on peut imaginer dans des bassins jurassiens, pré-alpins, appalachiens etc... les successions météorologiques qui feraient tomber de très grosses averses d'hiver, notamment en mars, sur des tranches anormales de neige. En fait, nous ne connaissons avec certitude qu'un seul exemple de grande crue catastrophique due à un hasard de ce genre. Ce fut celle de mars 1936 en Nouvelle-Angleterre, plus haut citée (1). Mais nos calculs ont montré que dans maintes grandes crues du Rhône à Lyon, de l'Ain, du Doubs, des rivières du Massif Central, de la Seine, attribuées avant tout à la fonte, celle-ci avait joué un rôle secondaire, ou minime. Lors du désastre de mars 1930, cependant, la neige a pu fournir 20 à 25 % des débits à l'Agout, mais moins de 15 et plutôt de 12 % pour l'ensemble du bassin du Tarn avant Montauban.

Notons que dans ces zones et dans les plaines, la fonte à elle seule ou presque seule, cause pas trop rarement des crues petites ou moyennes, pour lesquelles il suffit de 30, 40 ou 50 mm d'eau atmosphérique. Mais pour les inondations exceptionnelles qui exigent par exemple 100, 150 mm ou plus, la neige, sauf par exception, ne peut fournir qu'une fraction habituellement modérée du volume total.

c. — Plaines et petites montagnes d'Europe centrale

La fonte semble exercer une influence plus considérable dans les bassins de l'Elbe, de l'Oder, etc. et aussi dans la plaine de l'Ohio aux Etats Unis, parce qu'en ces régions à hivers très frais, la neige peut occuper plus d'étendue et sur une plus grande épaisseur que dans l'Europe occidentale. La grande crue générale de l'Ohio en février 1884 fut en partie nivale. Et celles de mars 1845 sur l'Elbe supérieure, de mars 1891 sur l'Oder et bien d'autres de ce genre semblent devoir à la neige une part non négligeable, sinon primordiale de leur gravité. Cependant, les inondations les plus remarquables de ces réseaux et surtout de celui de l'Oder ou de la haute Vistule ont pour causes des déluges pluviaux de plein été sans aucun apport de neige fondue.

3. — Crues des rivières nivales de plaine

Enfin, nous arrivons à de vastes contrées où la fonte des neiges produit presque tous les ans des crues imposantes ou même grandioses. Il s'agit des régions de climat continental froid où règne le régime nival de plaine : Russie, plaine canadienne, etc... La Volga monte à Kouibichef (Samara) de 10 à 12 mètres en moyenne, chaque année, sous l'influence primordiale de la fonte. Pourtant celle-ci ne libère sur les bassins, au dégel, que des lames d'eau relativement peu épaisses : 75 à 100 milli-

(1) En mars 1907, la grande crue (la seconde connue pour son maximum à Pittsburgh) de l'Ohio supérieur fut en partie nivale.

mètres par exemple, 120 à 150 dans les cas exceptionnels. Mais toute cette réserve fond d'un seul coup (nous voulons dire en peu de jours, parce qu'il n'y a aucune différenciation décisive dans le relief). C'est cette simultanéité relative, au rebours de l'échelonnement dans le temps constaté pour les bassins de haute montagne qui rend si brutales les crues russes, de dégel, et produirait les mêmes effets au Canada si des lacs de ce pays n'interceptaient point de façon très efficace les excédents de débits. D'ailleurs pour le Canada comme pour la Russie nous ignorons si les maxima exceptionnels, comme ceux de mai 1926 pour la Volga, ont dû leur gravité uniquement à la neige.

4. — *Rétention nivale*

L'examen de ces mécanismes confirme que, tout en n'admettant point une foule d'exagérations, nous ne nions point la production de crues nombreuses et, en certaines régions, très considérables par la fonte nivale jouant un rôle prépondérant, voire unique, ou subordonné.

Mais en contre-partie nous ne saurions trop insister sur le fait que le *facteur neige, supprime et réduit au moins autant de crues qu'il en occasionne* ou en augmente. Il agit de la sorte, par la rétention, surtout en hiver sur les hautes montagnes. Dans les grands bassins alpestres ou pyrénéens, les averses automnales ou hivernales les plus tièdes pour les vallées s'accompagnent de températures inférieures à zéro et de chutes nivales sur des fractions plus ou moins vastes (moitié, deux tiers, trois quarts des bassins totaux), au-dessus de 1200 à 1500 mètres. D'où suppression en ces zones de débits excédentaires qui seraient très dangereux puisqu'à de telles dates, le quotient d'écoulement atteint d'ordinaire 70 à 90 % pour les averses graves. On peut donc affirmer que maintes villes situées sur les rivières de haute montagne (par exemple Grenoble, Innsbruck, etc...) ne pourraient exister sans la rétention nivale exercée par ces reliefs.

Et même en plaine, des chutes nivales dans les secteurs froids des aires pluvieuses cycloniques peuvent en maintes circonstances réduire la gravité de la crue. Le désastre de mars 1936 sur l'Ohio à Pittsburgh a été euasé en petite partie par des fontes sur les Appalaches. Mais cet appoint a été au moins compensé, par une rétention nivale active sur la haute Monongahela et surtout sur le bassin supérieur de l'Allegheny River, où soufflaient les vents glacés et neigeux du Nord.

Notons que même en saison chaude et contrairement à une opinion aussi commune que peu réfléchie, la rétention nivale lors des averses atténue souvent les crues des rivières de haute montagne. Le fait est facile à comprendre pour qui observe que les grandes pluies de printemps ou d'été, dans les Pyrénées ou les Alpes par exemple surviennent en même temps que de sensibles refroidissements. Les températures réduites s'opposent plus ou moins complètement à la liquéfaction de la neige. Et ce dernier facteur n'a point ou n'a guère agi, malgré la renommée lors des inondations célèbres de mai 1856 sur le Rhône, de juin 1875 sur la Garonne, de Mai-juin 1917 sur le Pô, etc...

5. — *Allure et distribution du ruissellement lors des crues causées en partie par des fontes*

Enfin, nous mentionnerons de façon sommaire un aspect non négligeable de la question : l'allure des montées et des baisses lors de ces phénomènes. Selon certains auteurs, une couche nivale préexistante absorbant une partie de la pluie retient celle-ci quelque temps. Elle peut même annihiler le ruissellement si l'averse est faible par rapport à la couche. Si la pluie persiste et s'aggrave tandis que la température croît, la neige, à la longue, devient incapable de retenir l'apport atmosphérique liquide, et le ruissellement grave commence. Mais de toutes façons, l'inhibition dans le blanc feutrage retarde et répartit la crue sur un temps total plus long, ce qui atténue le débit de pointe (1). Le fait paraît nettement s'être produit en mars 1936 dans les Appalaches. Par exemple dans le bassin de l'Allegheny, les volumes moyens maxima de 24 heures ne dépassèrent pas 20 % du volume total, là où fondit beaucoup de neige. Là où celle-ci n'amena aucun renfort, mais aucun retard, la proportion journalière maxima atteignit jusqu'à 30 %. Et dans la mesure où des comparaisons peuvent se faire, un aplatissement analogue paraît avoir eu lieu en

(1) Cette atténuation n'a lieu que par rapport au débit que causerait une pluie liquide égale à l'averse réelle jointe au renfort de fusion, mais de toutes façons le maximum est renforcé par rapport aux valeurs qu'il présenterait si la pluie constatée agissait seule.

cette circonstance partout où la neige ajouta au ruissellement un contingent liquide notable.

Cependant, nous croyons qu'il serait excessif de poser ce principe en règle générale : les choses doivent dépendre de l'intensité, du réchauffement, de sa durée, de l'épaisseur et de la texture de la couche. Si la température et la durée pluviale sont suffisantes, l'absorption prolongée à un taux considérable devient à un moment donné impossible, l'éponge fond, cède à une vitesse accélérée et par conséquent, l'intensité ruisselante maximum doit être très accrue. Ce serait comme si d'innombrables mais infinis barrages se rompaient dans un court intervalle alors l'écoulement maximum devient égal et vraisemblablement supérieur à celui que créerait une pluie liquide égale à la pluie réelle et à l'apport de fusion.

Mais cette opinion aurait besoin d'être soit vérifiée, soit précisée par des observations, des études multipliées. Nous avons vu que ces recherches sont délicates. Nous espérons cependant, que dans les années à venir elles seront entreprises ou perfectionnées et menées à bien dans tous les bassins où des phénomènes caractéristiques et susceptibles de constatations assez sûres auront eu lieu.

Maurice PARDE.

submitted by

ROY E. LUNDQUIST, Hydrologist, U. S. Department of Commerce, Weather Bureau Cincinnati, Ohio.

DESCRIPTIVE AND QUANTITATIVE CLASSIFICATIONS
OF VARIOUS CONDITIONS
PERSISTING IN SNOW AND ICE DEPOSITS

The need has long been recognized for adopting some standard method of classifying or defining the various conditions which are known to persist in snow and ice deposits. Accordingly, the following suggested outline was first proposed by the author on September 14, 1942, to the U.S. Weather Bureau as a continuation project to supplement the official definitions and symbol classifications relating to hydrometeors. The major purpose of this proposal was to bring about some form of standardization in the various programs relating to snow research and observation by both professional and lay observers. If such a proposal could be decided upon, it was believed that snow and ice observations and reports made by various agencies could be understood more readily by every one because the definitions would be uniform and specific. Application of the descriptive terminology when understood in terms of quantitative definitions would assist in classifying data relative to the many problems which arise from snow and ice conditions as they affect skiing, snow removal, snow-melt contribution to streams, insulation for ground surfaces and crops, evaporation and sublimation, snow surveys or simple observations for climatological records, and many others.

In the absence of proper equipment and professional techniques in observation, the reporting of snow conditions by ordinary observers who have studied or been trained in these definitions could form the basis of important deductions about snow density and probable water content, which information is highly pertinent to flood forecasting as well as water supply forecasting and analysis.

The outline suggested herewith is an improved arrangement of a system of notation used in field investigations of snow conditions in the U.S. Weather Bureau Big Cottonwood, Utah, Special Snow Studies Program and in a cooperative Snow Stability and Safety Program at the ski area in Little Cottonwood, Utah, at elevations varying from 4500 to 10,000 feet m.s.l. Symbols and group numbers are assigned to conform with the desirable practice of using condensed methods of recording observations and for simplicity in the transmission of reports. For further simplicity a code could be worked up to satisfy most or all classifications. After working with a copy of the definitions at hand, an observer readily masters, almost from memory, the various conditions that persist from day to day in snow.

The resort to a quantitative definition of the various snow classifications in terms of hundredths of an inch or in wire gage is simply to tie down arbitrarily some definite points of change from one classification to another. The grading of snow structure according to grain size seems most logical. When the grains are fused or consolidated into crusts or solid ice the grain structure is no longer defined.

The degree of compactness of the snow or ice structure is most readily represented by density compared to an equal volume of water at 4° C, standard conditions. A more extensive field check of the density brackets was undertaken at Soda Springs, California, but final report has not been made. The assigned densities are therefore provisional and are based on numerous observations made in freshly dug snow trenches in snow covers varying from a few inches up to 100 inches in depth, and at various periods of the fall, winter and spring months in the Wasatch Mountains of Utah.

The quality definition shows presence of melt-water. The technique for determining snow quality has been presented by Walter T. Wilson, in his paper entitled 'Some Factors in Relating the Melting of Snow to its Causes' published on pp 33 to 41, Central Snow Conference Proceedings, Volume I, December, 1941. Quality is defined as the percentage of ice in the mixture, recognizing snow as a mixture of water and of ice particles.

AN OUTLINE OF DESCRIPTIVE AND QUANTITATIVE CLASSIFICATIONS OF VARIOUS CONDITIONS PERSISTING IN SNOW AND ICE DEPOSITS

(For Purposes of Uniform Observational Notation and Description in Connection with Flood Forecasts, Runoff from Snow Storage, Ski Reports, Snow Slide Hazards, Radiation Penetration Studies, Evaporation from Snow and Ice, and Other Snow Research.)

I. Type of snow or ice deposit

A. *Fresh Accumulations*

(or new top layer — same as U.S. Weather Bureau definitions for Hydrometeors)

(Classification)	Notation Symbol	
1. Snow Flakes or Snow Crystals		
2. Snow Grains		
3. Snow Pellets		
4. Sleet (hard pellets of ice)		
5. Hail	 Small Hail	 Hail
6. Rime	 soft	 hard
7. Glaze		
8. Frost		

B. Old Deposits

(in cumulative deposits each storm adds a distinctive layer to unmelted previous layers)

(Classification — figures or gage
refer to average diameter range)

Notation Symbol

1. Fine Powder Snow (flaky crystals or needles-grains generally indistinguishable)



2. Coarse Powder Snow (grains becoming noticeable, less than .02'')



3. Fine Granular Snow (grains .02'' to .05'', # 25 to # 17 Standard Wire Gage)



4. Medium Granular Snow (grains .05'' to .125'', # 16 to # 11 Standard Wire Gage)



5. Coarse Granular Snow (grains .125'' to .20'', # 10 to # 5 Standard Wire Gage)



6. Corn Snow (grains larger than .20'', # 5 Standard Wire Gage plus)



7. Soft Crust (icy granulated layer on snow surface or persisting between snow layers, or on top of ground)



8. Hard Crust (same as next above but frozen into a harder layer)



9. Consolidated Ice (grains completely fused or solid crystalization)



II. Degree of Cohesion or Compactness

(Note : Densities are based on equal volume of water at 4° C as 100 %)

A. Loose Snow (Densities up to 20 %)

1. Fluffy (unconsolidated — readily blown or drifted by wind)
2. Loose (grains distinguishable but lacking cohesion)
3. Initially Cohesive (becomes loose when disturbed by hand or shoveling)

B. Firm Snow (Densities 20 to 35 %)

1. Firm (readily excavated by hand — shovels out in light chunks)
2. Very Firm (resists excavation by hand — chunks are firmer and heavier)

C. Compact Snow (Densities 36 to 55 %)

1. Compact (requires hard tool to excavate by hand — shovels out in heavy chunks)
2. Very Compact (difficult for one person to bore with snow tube — chunks are very heavy and are difficult to break out when shoveling)

D. Hard Snow and Ice (Densities 56 to 92 % — density of solid ice at 0° C = 91.7 %)

1. Unusually Compact Snow (as may be packed down on highways, in snow slides, drifts, etc.)
2. Granulated Crusts or Honeycombed Ice.
3. Solid Ice (such as forms in or between snow layers at old crusts planes and at the ground surface, or on surfaces, edges, etc., of rivers, ponds, and lakes)

E. Slush (at least 50 % — generally more than 75 % density)

1. Firm Slush (ice or snow turned to a soft but cohesive mass)
2. Amorphous Slush (mixed snow and water, or slush and water lacking cohesion or unable to support a heavier object)

III. Quality

(Relative amount of melt-water present in snow)
(100 % quality shows no melt-water, thus where temperatures below freezing are indicated in the snow, crystalization is complete and quality is 100 % or more)

A. Dry Snow (100 % quality or more — temperature in snow below freezing)

1. Fresh Snow will not pack in hand — will not support travel without skis or snow-shoes.
2. Old Snow is crumbly or packs with difficulty.

B. Moist Snow (Quality 90 to 99 % — temperature in snow at or very near 32° F or 0° C). (Snow packs readily when pressed in hand)

C. Wet Snow (Quality 75 to 89 % — temperature in snow at or very near 32° F or 0° C). (Snow drips water when pressed)

D. Dripping-Wet Snow (Quality less than 75 % — temperature 32° F or 0° C or slightly higher imparted by temperature of melt-water). (Snow drips water without pressure)

Note : The above definitions are based on field examination and tests from data recorded at the time of excavating numerous freshly dug snow pits in shallow and deep snow up to 100 inches.

THE FORMATION OF ICE CRYSTALS IN THE LABORATORY AND IN THE ATMOSPHERE

by

VINCENT J. SCHAEFER

General Electric Research Laboratory, Schenectady, New York

Paper prepared for presentation at the meeting of the international Commission for Snow and Glaciers as part of the Eighth General Assembly of the International Union of Geodesy and Geophysics to be held at Oslo Norway, August 27 and 28, 1948. The subject matter included in this paper was presented at the General Meeting of the American Chemical Society held in Chicago, Illinois April 23, 1948 as part of the Aerosol Symposium of the Division of Physical and Inorganic Chemistry and the 1948 Annual Meeting of the American Geophysical Union held in Washington, D. C. in April, 1948 under the sponsorship of the Division of Hydrology.

During the past seven years, various phases of snow studies have been carried out by the writer and several of his associates of the General Electric Research Laboratory at Schenectady, New York.

The earlier work was conducted in a snow laboratory on a slope of the Mohawk Valley in eastern New York. Subsequent studies starting in 1943 and continuing at the present time were also conducted at the Research Laboratory at Schenectady, New York and the Mt. Washington Observatory in the state of New Hampshire.

The work has included such studies as the electrical properties of snow, the physical effects causing precipitation static on aircraft radio, various techniques for making snow crystals replicas, studies of the changes in snow on the ground, the physical processes related to the deposition of ice on aircraft surfaces, a study of supercooled water droplets, both in bulk phase and as an aerosol, methods for detecting sublimation nuclei in the free atmosphere, and finally, various phases of cloud physics, particularly « seeding » techniques which may be used to « trigger off » unstable cloud systems.

Space nor time permits more than brief descriptions of the various subjects which have been or are being studied at the present time. References are made, however, to papers which are available in the literature or as reports issued by our laboratory which may be used to obtain more detailed information on the various subjects discussed.

The Electrical Properties of Snow

Early in 1943 several of us in the Research Laboratory were asked to make a study of some of the causes of precipitation static. This is a trouble experienced when aircraft fly through various types of precipitation. Under such conditions and, particularly when the precipitation is in the form of snow, much static noise develops on the aircraft radio. Under severe conditions, the radio becomes useless because of the high noise level, and at times St. Elmo's fire becomes prevalent on the wing and propeller tips and other parts of the airplane.

To better understand certain phases of the phenomenon, the writer made a study of the electrical characteristics of snow particles and snow storms. The general results of this research have been published. (1)

In brief, the studies indicated that one of the main causes of static noise is the charging effect produced when snow particles hit a plane and are shattered, the fragments rubbing and bouncing across the metal surface thus serving as units of a frictional electrical generating system. The studies showed that a single snow crystal may be broken into thousands of fragments. As the plane becomes charged, it finally reaches an unstable condition at which time an electrical discharge occurs which may assume the form of St. Elmo's fire as previously mentioned. These electrical discharges account for the radio noise.

Among the interesting effects observed during the study of the electrical properties of snow storms was that two types of snow storms could be identified in terms of their electrical properties. These were termed Cross Current and Positive Current storms, the evidence indicating that the former were storms having clouds with a considerable amount of convective activity while the latter storms apparently consist of relatively stable clouds.



FIG. 1. — Satisfactory exposure for point collector to record atmospheric electricity.

The electrical activity during storms is easily measured by mounting a well insulated collector such as a needle in a well exposed position as shown in Fig. 1. If the needle is connected by shielded cable to a recording microammeter and then to ground, records such as shown in Fig. 2a and 2b are obtained in practically all snow storms. After studying many storms, Table I was developed to show the general characteristics of the storms producing these two interesting types of atmospheric electricity.

The formation of ice on Aircraft

One of the interesting developments following our study of precipitation static at the Mt. Washington Observatory was a shift of our interest from that subject to one of even more importance — the icing of aircraft.

Since Mt. Washington is noted for having « the worst weather in the world », it was found that the summit was an ideal spot to study some of the basic problems related to the icing of surfaces from supercooled clouds. Studies at the Observa-

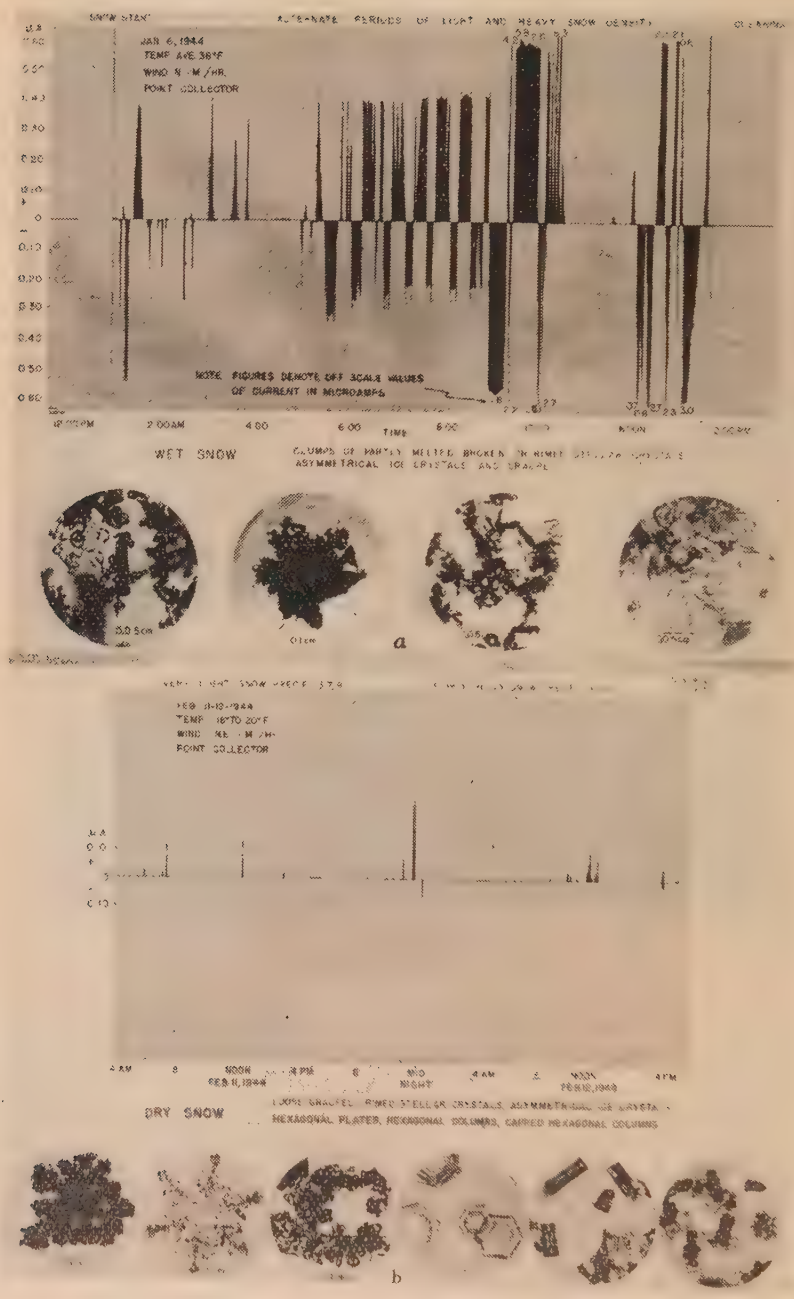


FIG. 2. — Typical atmospheric electricity from positive current and cross current snow storms.

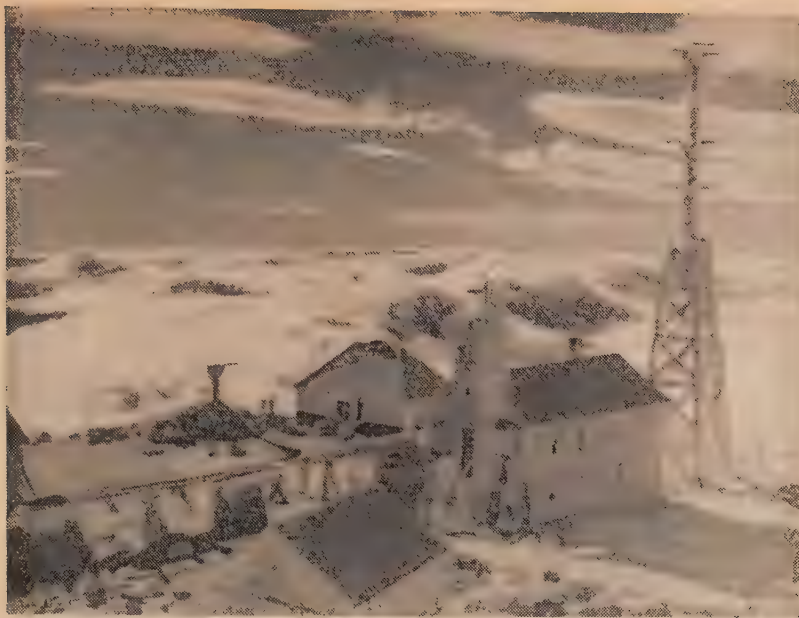


FIG. 3. — VIEW OF Mt. WASHINGTON OBSERVATORY and Supercooled clouds.

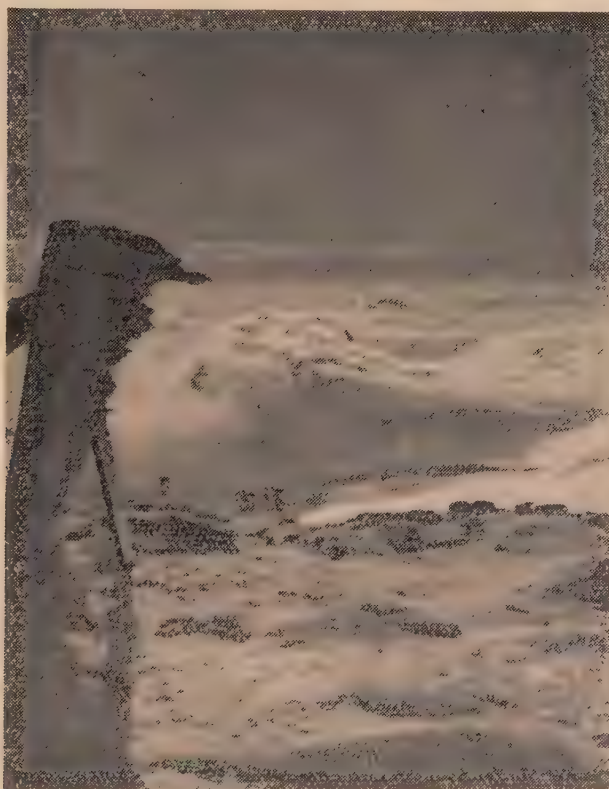


FIG. 4. — View of Snow clouds from summit of Mt Washington, New Hampshire.

tory have been carried out each season since 1943 in an effort to better understand some of the fundamental phenomena. Much of the results of this work has been published in reports (2), most of which are available from either the U.S. Air Forces or our laboratory.

It was the writer's visits to Mt. Washington and many observations made on the arduous climbs and descents of the mountain during icing and snow storms that eventually led to the studies resulting in cloud modification techniques. A view of the Observatory after a mild icing storm is shown in Fig. 3. The clouds above and below the summit are all in a supercooled condition. Under somewhat similar cloud conditions but later in the season, Fig. 4 shows a relatively thin orographic cloud system which shed snow continuously for several days. A consideration of these observations and many others laid the groundwork for subsequent discoveries.

The Classification of Snow Particles

Any fundamental work with snow storms and snow crystals immediately points out the necessity of having a simple classification chart to record the crystal forms occurring at a specific time from a storm.

TYPES OF FROZEN PRECIPITATION					
CODE	GRAPHIC SYMBOL	TYPICAL FORMS			TYPE
1					PLATES
2					STELLARS
3					COLUMNS
4					NEEDLES
5					SPATIAL DENDRITES
6					CAPPED COLUMNS
7					IRREGULAR CRYSTALS
8					GRAUPEL
9					SLEET
0					HAIL

PROPOSED CLASSIFICATION OF SOLID PRECIPITATION
INTERNATIONAL COMMISSION ON SNOW AND ICE, I.C.S.I. COMMITTEE ON SNOW CLASSIFICATION

FIG. 5. — Classification chart of various forms of frozen precipitation. (revised)

Other investigators in this field (3) have suggested classification methods, but a survey indicated that they were either more complicate than necessary or else failed to include one or more important types. The classification chart shown in Fig. 5 was prepared by the writer in 1943 for a group of cooperative observers who assisted the writer in making a study of the distribution of crystal forms which occur during widespread snow storms in the northeastern part of the United States. This has been revised several times as the result of experience in the field, but in its present form, has now been successfully used for more than three years.

In order to describe the crystal forms observed at a particular time in a snow fall, the code number related to the crystal type is preceded by the letter P. If the crystal is coated with cloud particles so that it has a frosted appearance, the letter R following the code number denotes that condition. Thus the designation P 1 R, 6, 7 indicates that at the observation time stellar crystals coated with rime, ice needles and asymmetrical crystals were falling from the sky.

Greater detail may be included by affixing a subscript number indicating by tenths the relative quantities of each type at the observation time. The photomicrographs used to illustrate the chart show some of the variations which occur with each general type of snow particle. Table II indicates the general features

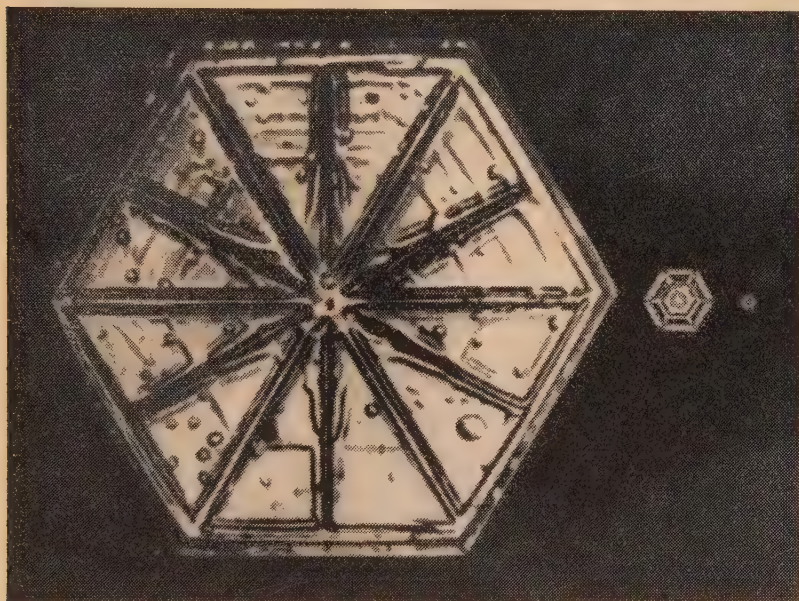


FIG. 6. — Relative sizes of hexagonal plate type of snow crystals.

and relative sizes of the nine different forms of precipitation. Fig. 6 shows the range in size which is commonly observed with a typical crystal form — the hexagonal plate. The larger one comes from relatively low level clouds containing moderate quantities of moisture. Such clouds are generally below the 10,000 foot level. The crystal shown in the middle of the figure is common to cirrus type clouds and may fall from an altitude as great as 40,000 feet. The smaller crystal is the type known as diamond dust which commonly forms when the air is supersaturated with respect to ice. Such conditions generally happen with strong temperature inversions accompanied by cloudless skies, the crystals coming from a region less than 1,000 feet above the ground.

Cold Chamber Studies

The basic experiment (4) in which a supercooled cloud of liquid water droplets may be converted to a mass of snow crystals is a spectacular one which may be easily reproduced with very simple apparatus.

The essential articles for the experiment are an open topped, well insulated chamber having a volume of about $1/20$ cubic meter, e. g., 48 cm long, 24 cm wide, 48 cm deep, which may be cooled to a temperature of -10 to -20°C . The walls and floor of the chamber should be painted dull black or lined with black velvet, cloth, or cardboard.

Illumination is simple but important. For best results, the light should produce a brilliant beam and is most satisfactory when mounted above the chamber and directed into it at an angle of 45 degrees from the horizontal.

A supercooled cloud is produced by placing a moist cloth in the cold air of the chamber. The moisture condenses into small water droplets, generally in the size range of $10 - 20$ microns diameter. Another simple method for forming the cloud is to merely breathe into the cold air. Several breaths will produce a cloud having a liquid water content equivalent to 1 gram per cubic meter. A chamber with a cloud in it is shown in Fig. 7.

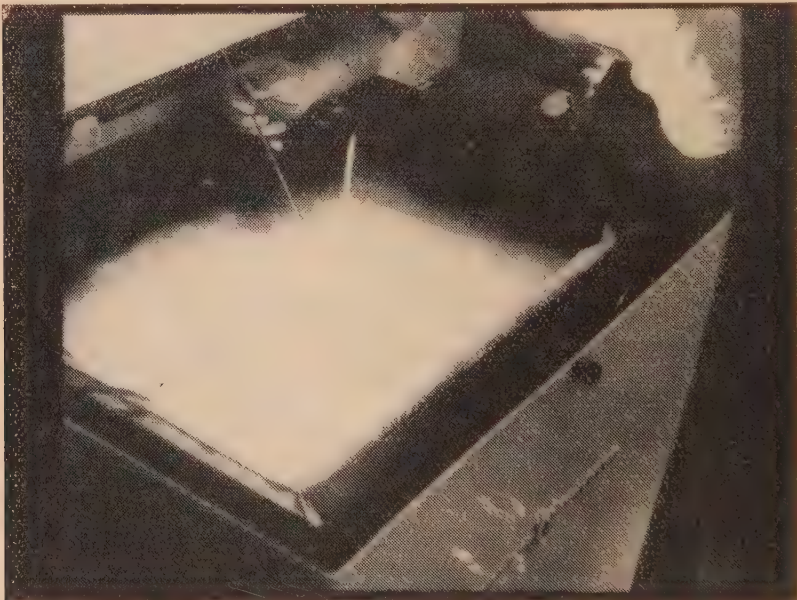


FIG. 7. — View of cold chamber containing supercooled cloud.

Under ordinary conditions, few crystals will be seen in the chamber. A single crystal may easily be seen among the water droplets since it grows rapidly, scintillating in the light. They are most easily seen when viewed at right angles to the axis of the beam since much of the light from them is specular reflection. This is particularly true of hexagonal plates since they become oriented with their flat surfaces horizontal.

It can easily be shown that most of the condensed cloud formed in the chamber is supercooled. If a fine wire is rotated at 180 r.p.m. in the chamber, it becomes coated with ice. If left alone, the cloud slowly disappears by condensing onto the cold walls of the chamber.

Within a minute after the supercooled cloud has been formed, the crystals which develop from sublimation nuclei present in the air grow and fall to the floor of the chamber. The supercooled cloud may be suddenly converted to snow crystals by briefly introducing an object colder than -39°C . This may be a needle

cooled in liquid air or a tiny particle of dry ice. A single passage of such a cold object produces an effect like Fig. 8 which quickly changes to a mass of twinkling crystals like Fig. 9. These crystals range in size from 10 to 200 microns. By the time they reach this latter size, the falling velocity is of the order of 1 cm sec^{-1} so that it is difficult to keep the particles from « snowing out ». Typical crystals produced in this manner are shown in Fig. 10.

One of the most convenient ways to convert supercooled clouds to ice crystals is by dry ice. Dry ice is easily produced by permitting liquid carbon dioxide to expand into the air. A very convenient source of dry ice for experimental studies is the small 10 gram shells, such as used in soda fountains and in various home

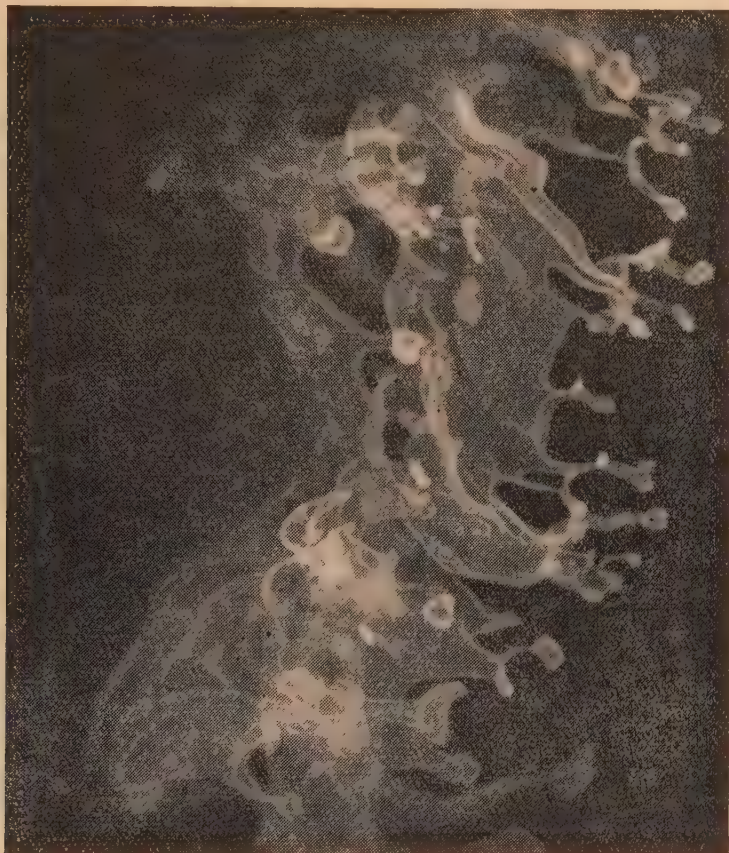


FIG. 8. — Trail of ice crystals produced in cold chamber by seeding with needle cooled below -39°C .

appliances. A simple valve for controlling the opening of such containers is available as part of a small model airplane carbon dioxide engine. By valving the container held in an inverted position above the cold chamber, tiny bits of dry ice will be seen to fall through the cloud leaving behind tiny white « contrails ». These rapidly grow until the cloud is completely replaced by a mass of twinkling ice crystals.

Several very interesting phenomena may be seen in this experiment. The cold chamber, when operated with the lid removed, has a temperature gradient with the air warmest at top, colder below. This temperature inversion produces a very stable air condition in the chamber which resists convection and turbulence. Thus

when seeded with dry ice or a cold needle, the heat released by the crystallization process is not sufficient to overcome the inversion stability. If the ice crystal region is observed carefully, a dark area quickly develops at the interface between them and the supercooled cloud. This is caused by the fact that the ice crystals grow at the expense of the water droplets which evaporate, the ice crystals growing from the condensation of gaseous water vapor since the vapor pressure of ice is below that of water at all temperatures below 0°C . The dark area often grows to a width of 1.5 to 2 cm in a few seconds showing in a striking manner the phenomenon of diffusion. Another interesting and unusual effect is observed whenever the concentration of ice crystals in the chamber exceeds about 500 /ccm. In addition to the glint from the crystals as they tumble about in the chamber, a background twinkling



FIG. 9. — Twinkling crystals formed within less than 1 minute after seeding of supercooled cloud.

is observable. This is due to Brownian motion and is an elegant way to see this interesting phenomenon.

Many other experiments could be described which may be carried out in the cold chamber. These must be deferred at this time, but it should be emphasized that the cold chamber as described is so simple to use that it should become a familiar research tool and demonstration unit for those working in meteorology and other related physical studies.

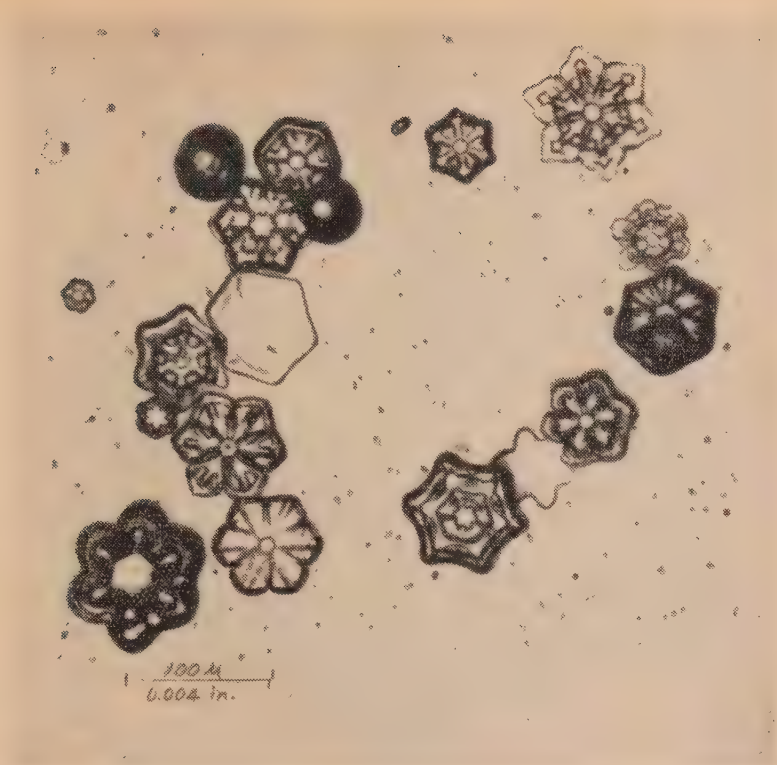


FIG. 10. — Typical photomicrograph of snow crystals produced in cold chamber by dry ice seeding.

The Preparation of Snow Crystal Replicas

Excellent replicas of snow crystals may be prepared by a very simple method which has been described elsewhere (5).

It consists briefly of using a dilute solution of a resin in a solvent which does not dissolve ice. An excellent solution consists of one percent polyvinyl formal dissolved in ethylene dichloride. When cooled below 0°C , it may be used to produce exact replicas of snow crystals by letting natural crystals fall on a collecting sheet of black velvet cloth or paper and then placing them in a small drop of solution previously placed on a piece of cold glass or slightly absorbent black cardboard. A variation of the technique is to coat a cold glass plate or black cardboard sheet with a thin film of the replica solution which is then exposed to the falling snow crystals for a specific time period ranging from five to sixty seconds depending on the density of the snow fall. This produces a permanent record of the snow fall at that particular time in terms of snow crystal type, concentration of crystals, and other factors of importance to obtaining a better understanding of the physical features of such storms.

The technique is so easy that relatively untrained observers can obtain excellent results by observing a few basic rules. The method also lends itself admirably to sampling from planes, at remote stations, and by automatic exposure instruments. The samples thus obtained may then be studied at leisure in a warm laboratory. Photomicrographs of replicas are shown in Fig. 11.



FIG. 11. — Photomicrograph of snow crystal replica.

The Preparation of Replicas of Snow on the Ground

In addition to the various replica techniques previously described, there is another which may be used to advantage in preparing replicas of snow on the ground. For this purpose, a more concentrated solution of Formvar should be used having a concentration of between three and five percent.

To illustrate one of the possibilities, replicas were prepared of a cross section of snow on the ground as shown in Fig. 12. To prepare these records, a five centimeter strip of three percent solution was spread lengthwise on a piece of window glass 24 inches long. A vertical face was cut into a fresh blanket of snow and the coated glass pressed firmly against it, withdrawn, and then placed in a well ventilated but sheltered location to permit the solvent to evaporate. Until all traces of the solvent has evaporated, the replica must be kept at a temperature colder than 0°C . The best replicas are obtained when the snow particles are permitted to sublime. If it is desirable to hasten the process, however, the sample may be warmed as soon as the solvent is gone which may take from 20 to 60 minutes, depending on the ventilation and the temperature of the replica sample. After obtaining a cross section replica as described, subsequent samples may be obtained by successive applications of the solution side by side until the surface area of the backing is exhausted.

The sample shown in the photograph illustrates a vertical shrinkage of the snow layer during a period of five successive days during which time the air temperature remained continuously below 0°C . Although the shrinkage amounted to more than fifty percent in snow thickness, the density increased from less than 0.10 to more

than 0.20 so that the liquid water content remained nearly the same. Examination of the structure of the snow particles on the fifth day showed that although most of the original snow consisted of perfect crystals of stellar, hexagonal plate, and column types, a major change occurred during the five day period producing granular snow approaching the rounded firm forms.

Subsequently, as shown by the photograph, several additional light snow falls occurred which were also recorded by the replica technique.

It is hoped that eventually a snow replica solution may be developed employing the solventless type of plastic in which several monomers are catalyzed by an oxi-

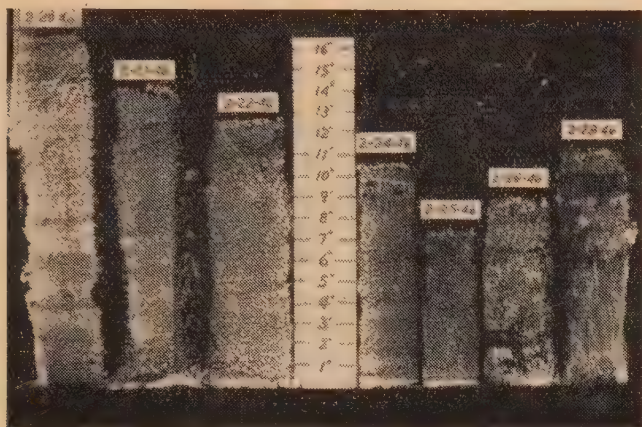


FIG. 12. — Replica of cross section of snow deposits on ground.

dizing agent to form a true three dimensional structure. Until such a material becomes available which employs materials which do not modify ice, satisfactory replicas may be prepared as described particularly when good ventilation is provided.

The Detection of Ice Nuclei in the Free Air

The random occurrence of supercooled clouds in the free atmosphere is a well known fact. The reason for this seemingly haphazard phenomenon is probably closely related to the concentration of sublimation nuclei in the particular air mass when clouds occur.

A better knowledge of sublimation nuclei is needed, particularly of particles other than ice which lead to the formation of ice crystals in air supersaturated with respect to ice. Unfortunately, such particles are generally of microscopic or sub-microscopic size and, in addition, include only a small fraction of the dust in a given sample of air.

It is important to differentiate between *condensation* and *sublimation* nuclei. The former are always much commoner than the latter. As will be shown in a subsequent chapter, a third type of particle, the *freezing* nucleus is also of importance in this connection.

Three methods have been devised by the writer for detecting sublimation nuclei in the clear air of the atmosphere (6).

The most satisfactory of these methods is to use the cold chamber of the type previously described. A sample of the surrounding air is introduced into the chamber by simply removing the air within it by employing a suction fan. Unless the air which is introduced has a very low humidity or was nearly as cold as that in the chamber, condensation will occur. By illuminating this newly formed cloud with a strong beam of light, every active sublimation particle which was present will appear as a twinkling snow crystal, easily discernible among the surrounding super-

cooled cloud droplets. The temperature of the air in the cold chamber will have a direct relationship to the number of particles likely to be observed.

Observations made during 1947 and 1948 on the summit of Mt. Washington, New Hampshire (6,288 feet) and in the Mohawk Valley show the concentration of sublimation nuclei at a cold chamber temperature of -20°C to vary from none observable to values up to ten per cubic centimeter. The type of crystal which forms under such conditions probably bears a definite relationship to the substance serving as the active nucleus. These crystals may be examined by direct examination under a microscope kept in the chamber, by preparing plastic replicas, or by permitting the crystals to seed a thin supercooled water film.

Certain « soap » bubble solutions serve admirably for this latter purpose. Most solutions work except those containing large amounts of glycerine or similar freezing point depressants. A film is formed on a ring five to ten centimeters in diameter



FIG. 13. — Supercooled bubble in which ice crystals are growing.

and then held in the cold air of the chamber. The film supercools in about a second. If visible crystals are in the air, the number of separate crystals which develop and grow in the film appear to be in direct proportion to that concentration. At least four types of crystals grow in the film. A six membered stellar form appears when the crystals are hexagonal plates. A symmetrical cross or α -shaped crystal appears when hexagonal columns occur. An irregular shaped crystal predominates when asymmetric crystals are present. It should be remembered that there may also be crystals developing from freezing nuclei within the solution. Whenever this occurs, the crystals tend to have frost-like forms with a tendency to be curved. There

is a time lag also before such crystals appear as compared to the effects seen when sublimation nuclei or ice crystals are present.

A variation of this method is to blow bubbles in the free air when its temperature is below freezing. When this is done, crystals will sometimes develop in the bubble. Occasionally a single crystal will grow completely around a ten inch diameter bubble. Fig. 13 illustrates an effect commonly seen. Table III and IV show typical data indicating the variations which have been observed during the past winter.

Permanent replicas of such effects may be prepared in the following manner. On a frame of convenient size, such as 9×12 cm, a thin film of cellulose acetate is stretched and cemented. This film is then coated with a thin layer of three to five percent of polyvinyl alcohol dissolved in water. This solution readily supercools and when seeded with sublimation nuclei or ice crystals will produce effects identical with those already described for the soapfilm. Fig. 14 shows two stages of the

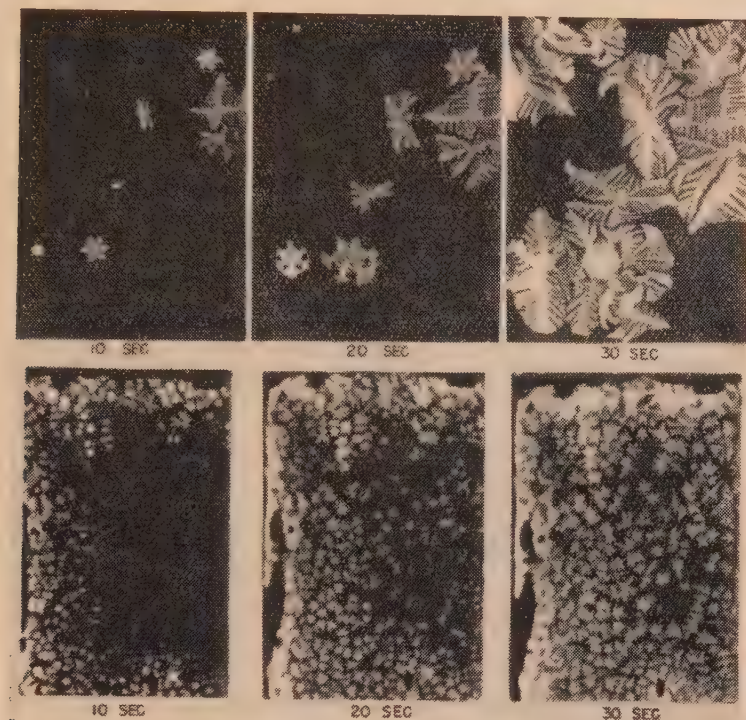


FIG. 14. — Stages in growth of ice crystals in supercooled 3 per cent solution of polyvinyl alcohol
Top, effect produced by 10 nuclei bottom effect produced by seeding.

growth of ice crystals in such films. If, after the crystallization is completed, the film is kept below freezing until the ice disappears by sublimation, a permanent replica remains. This was formed by the polyvinyl alcohol being precipitated at the interfaces between the ice crystals.

The Effectiveness of certain Dusts as Sublimation Nuclei

It is obvious that there are physical processes in the natural atmosphere which lead to the formation of snow at temperatures only a few degrees below 0° C. In the past, it has been suggested (7) that graphite and quartz dust carried into the atmosphere were the active sublimation nuclei for ice crystals. The hexagonal structure of these materials in their crystalline form were thought to be of importance

in their activity. So far as I am aware, no information has been published of any verification of these ideas by laboratory experiments or observations of nuclei of actual crystals.

It was found (8) by Dr. Vonnegut of our laboratory that small particles of silver iodide were quite effective as ice nuclei. This material has a rather high temperature coefficient, the warmest temperature at which any particles are active being about -5°C with a much greater activity at -10°C .

During the past year, I have found a number of naturally occurring dusts which serve as effective sublimation nuclei. Among these are volcanic dusts, silicates, and clays. In most instances, these materials have been selected from extensive natural deposits in arid or semi-arid regions noted for their dust storm activities. Each material has its own temperature range of activity for producing effective nuclei. For example, a sample of loess from an ancient lake bed near Richland, Washington when formed as an airborne dust shows the presence of a small percentage

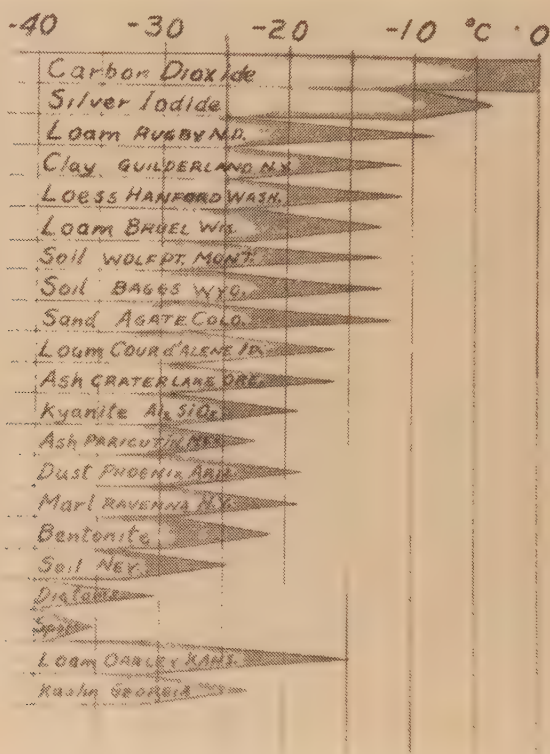


FIG. 15. — The dependence on temperature of certain foreign particles which serve as ice nuclei.

Full width indicates temperature at which all particles in sample are active; pointed end indicates temperature where a few particles are active.

NOTE. Carbon dioxide is effective due to temperature of -18.5°C not chemical composition.

of particles that form snow crystals in saturated air at a temperature of -12°C . At -24°C , all of the particles are « active » nuclei. Another dust-like soil from a desert in Nevada does not show any activity until the temperature reaches -24°C and does not reach one hundred percent activity until the temperature is -33°C . The temperature-activity relationships of a number of such natural dusts is shown in Fig 15 and several typical snow crystals grown on such foreign nuclei are shown in Fig. 16.

It is believed that the mechanism responsible for the activity of such particles

is the fact that the surface of the particle is conducive to the formation of frost. As soon as a frost deposit occurs on the surface of the particle, growth continues as though the particle was made up only of ice. Dust particles larger than 1 micron generally produce asymmetrical crystals with irregular plate-like growths, although smaller particles often lead to the development of very symmetrical ice crystals.

Since these dusts are active in the temperature range where snow is commonly observed, it is quite likely that they are responsible for the initiation of many snow storms. It is unlikely, however, that such materials would be so concentrated that they could provide the tremendous quantities of nuclei necessary to keep a snow storm going. It has been my observation that the majority of snow particles from storms coming from relatively low clouds are asymmetrical crystals which seem to be growing on fragments of other crystals. This strongly suggests that snow storms might often be initiated by foreign particle sublimation nuclei in the early stage of development of the storm. These particles grow big enough and have a sufficient concentration that for a short time they collide with one another. The resulting

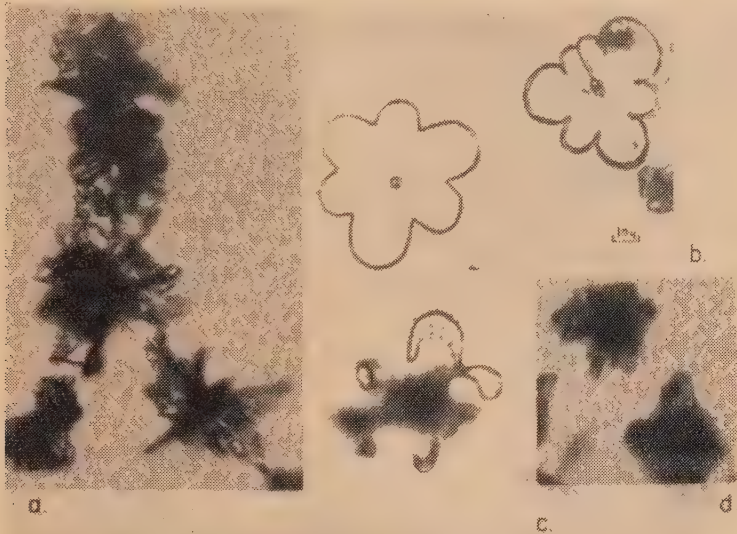


FIG. 16. — Typical snow crystals formed on foreign particle nuclei.

fracture from collision and abrasion and the electrical effects which result would then produce large numbers of tiny ice fragments, each of which would serve as an effective nucleus up to the freezing temperature. This mechanism is thus a chain reaction of a type somewhat comparable to that described by Langmuir (9) for the development of precipitation in warm clouds.

The Modification of Snow Crystal Forms by Vapors

The variations in snow crystal forms shown in Fig 5 are generally believed to be due to variations in temperature and moisture supply. It has been shown by Nakaya (10) that most of these variations in crystal form may be duplicated in a laboratory cold chamber by carefully controlling the temperature and moisture supply as a crystal grows on a supporting fiber.

Work in our laboratory (11) shows that modifications to the crystal form may also be produced by introducing certain organic vapors into the chamber in which the crystals grow. Unlike Nakaya's experimental method in which single crystals are grown, our experiments generally produce a minimum of 10^7 crystals.

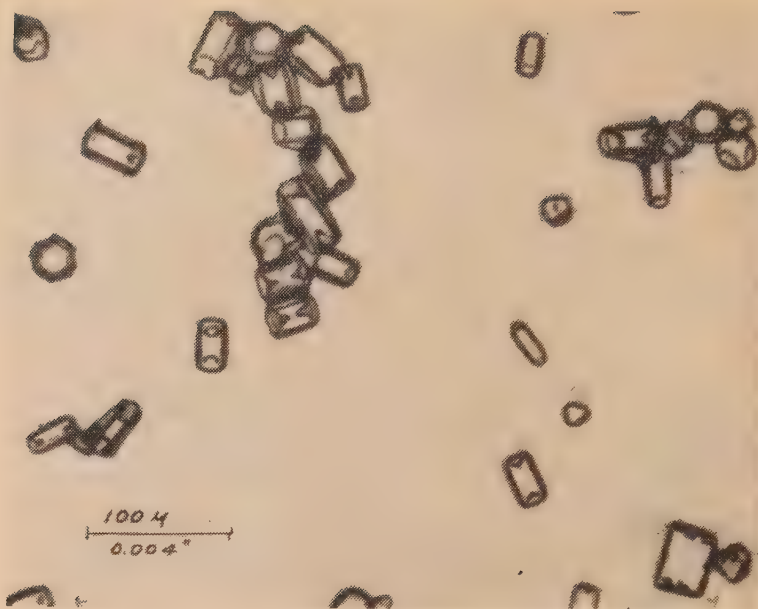


FIG. 17. — The effect of acetone vapor on the crystal habit of ice. The normal plate structure was shifted to a columnar form by 1 molecule of acetone to 10 of water vapor.

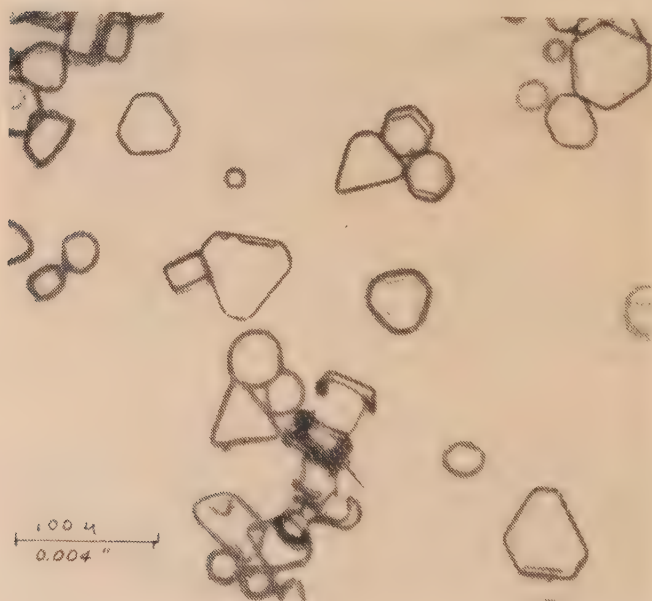


FIG. 18. — The effect of nitric acid vapor in air of cold chamber on the crystal habit of ice.

A typical experiment employs air having a mean temperature of -15°C . Moisture is added to the air by the evaporation of water from a porous porcelain container or by use of a nebulizer (a small aspirator constructed within an outer glass container which directs an atomized water stream against the surface of the outer wall, the finer particles and accompanying saturated air emerging as a stream of water condensate from a tube on the top of the nebulizer).

Under normal conditions when the supercooled cloud is « seeded » with dry ice or when adiabatic expansion cools a small sample of the air below -39°C , the ice crystals which form in ordinary laboratory air are simple hexagonal plates. When however, a small amount of certain organic chemicals are introduced into the air, either by vaporization or by use of the nebulizer, the crystals which form are modified. For example, as shown in Fig. 17, acetone in concentrations equivalent to one molecule of acetone to 100 molecules of water causes all of the crystals to become either hexagonal columns or a peculiar combination of plate and column. On the other hand, the introduction of the vapor of nitric acid or nitrous oxide, Fig. 18, tends to emphasize the trigonal symmetry of the water molecule so that the majority



FIG. 19. — The effective of acetic acid vapor on the crystal habit of ice.

of the crystals are triangular with some of them being almost perfect isosceles triangles with sharp corners bearing no suggestion of the hexagonal pattern so common to crystals formed in « normal » air.

The use of acetic acid, Fig. 19, encourages the development of large stellar-like crystals containing intricate structural designs. Larger molecules of the fatty acid series having an appreciable vapor pressure, such as valeric and caprylic acid, produce the columnar form of prism similar to that observed with acetone.

The aldehydes also modify the ice crystal structure, formaldehyde causing rather short columns while heptaldehyde produces a peculiar crystal somewhat akin to the capped hexagonal column.

Other chemical types which produce definite modifications are dioctyl sodium sulfo succinate (OT 100), heptyl and octyl alcohol (methyl and ethyl are not effective), certain of the silicones, notably dimethyl dichloro silane, pentamethylamino-methylidisiloxane, and the ketones.

A preliminary consideration of the results shows that the most effective chemicals for producing modification are those which have strong polar properties. These are generally surface active substances and presumably produce the modification

by becoming preferentially adsorbed on certain faces of the crystals. This restricts growth in some directions, allowing more growth on others.

The study up to the present time indicates that the modification of the crystal is a continuing phenomenon and not one which is directly controlled by the crystal habit of the nucleus. By producing a batch of crystals by the adiabatic expansion of air containing acetone vapor, then permitting them to grow in saturated air under normal conditions, a typical crop of hexagonal plates was produced.

This experiment was made by placing a tiny wad of filter paper soaked in acetone into a « pop » chamber having a capacity of 0.25 ccm. Saturated air at 25° C was introduced into this chamber under a pressure of 30 pounds per square inch. By the sudden opening of the chamber using a high speed release system, the air expanded and cooled below the critical temperature of — 39° C causing the spontaneous formation of ice crystals in great numbers. These grew at the expense of a supercooled cloud in the cold chamber. Instead of forming hexagonal columns as would be the case if the original crystal habit was dominant, the crystals which formed were of the « normal » variety of hexagonal platelets.

Work in this field is continuing since many fundamental phenomena in the field of crystallography may be studied with great ease with the techniques which have been developed.

Nomenclature in Experimental Meteorology

As the new science of experimental meteorology develops, especially as applied to cloud seeding and modification, certain experimental procedures are necessarily so often described that it becomes worthwhile to introduce new terms to simplify description and reduce the chance of ambiguity.

Since clouds occur in air which may be warmer or colder than 0° C, it was decided by our group to follow a suggestion of Langmuir's and adopt the following terminology.

The term a *cold cloud* shall be used to describe a cloud of water droplets which is supercooled and all of the cloud is in air which is colder than 0° C.

A *warm cloud* consists of a cloud of water droplets existing in air of which all parts are warmer than 0° C.

A *cool cloud* shall be the term used to describe a cloud which combines the properties of both *warm* and *cold* clouds; that is, the cloud exists in both above and below freezing air, with the freezing level extending through some part of it.

Other terms are being devised to describe some of the observable features peculiar to cloud modification studies. These will be described in greater detail in a later paper.

The Production of Snow in Supercooled Clouds

As a result of the experimental studies made in the cold chamber during the summer of 1946 (4), which has already been mentioned, the first experiment in the free atmosphere carried out by the writer took place on November 13, 1946. On that day a four mile supercooled alto stratus cloud having a temperature of — 18.5° C and containing no trace of ice crystals was « seeded » with six pounds of dry ice by flying through it in a small Fairchild monoplane at an altitude of 14,000 feet. Half of the fragments were put out from a mechanical dispenser operating on the floor of the plane, the rest scattered by merely letting the slip stream suck the fragments from cardboard containers held out of the open window. Fig. 20 shows the cloud before and after seeding.

Since that time many experiments have been carried out by the writer, by members of « Project Cirrus » — a joint Army-Navy-Air Forces-General Electric weather research program —, and by many other investigators. Some of these results may be studied by referring to detailed reports and papers (12). Positive results are always observed when supercooled stratiform clouds are seeded with dry ice at the rate of about one pound per mile. The magnitude of the effect is dependent on the stability of the air, the thickness of the cloud, and whether or not the air flow into the region is of a convergent nature. If convergence is present, it is quite possible that the « triggering effect » produced by dry ice seeding would result in a self-perpetuating storm spreading out from the area originally modified. Similarly cumulus clouds may be profoundly modified with dry ice whenever the cloud contains supercooled water droplets.

Laboratory experiments indicate that at least 10^{16} ice nuclei may be produced from one gram of dry ice when it is introduced into a supercooled cloud. If all of these particles could be supplied with moisture and allowed to grow to a small snow crystal, some 300,000 tons of snow would result. It is, of course, impossible to achieve such results under natural conditions since there is not enough moisture available in the area that can be affected. The important fact to recognize in this respect is that by using dry ice it is quite feasible to produce nuclei on the tremendous scale that is characteristic of natural weather phenomena. For example, the number of snow crystals required to produce one foot of snow over an area of 10,000 square miles is approximately 1×10^8 . That number of ice nuclei could easily be produced with two pounds of dry ice. Since it would not be feasible to spread two pounds of dry ice over such a large area, it would, however be practical to use 200 pounds at the rate of $\frac{1}{2}$ pound per mile and exert a large influence over an area of 10,000 square miles.



FIG. 20. — View of first cloud seeded by V.J. Schaefer Nov. 13 1946, using 6 pounds of crushed dry ice.

Under actual seeding operations, it has been found that about one pound of granulated dry ice will convert five cubic miles of supercooled stratus clouds to snow in about 30 minutes when the dry ice is introduced from a two inch pipe in a plane flying at the top of a cloud having a vertical thickness of a mile. Dry ice pellets in the size range 1-10 mm in diameter are used so that the crystals are produced within a minute or so along a plane extending from top to bottom of the cloud. The heat generated in the crystallization process ranges from 0.5 to 1°C in such a cloud causing turbulence which spreads the crystals upward and outward. Thus it is a common observation to see the affected area expand as shown in Fig. 21 at a rate of nearly five miles an hour. Fig. 22 is a photo of such an area.

So many crystals are produced when dry ice is dispensed at the rate of one pound per mile that the affected region often shows a persistent, stable snow crystal cloud remaining in the modified region after all of the adjoining clouds have disappeared by precipitation. Such an effect is recognized as « overseeding » but is only experienced with very stable clouds, thin clouds, or under conditions free of convergence.

Such effects would not be expected with cumulus clouds unless they were dissipating or when large quantities of dry ice are used.

Some evidence has been seen that extensive cumulus activity may be generated

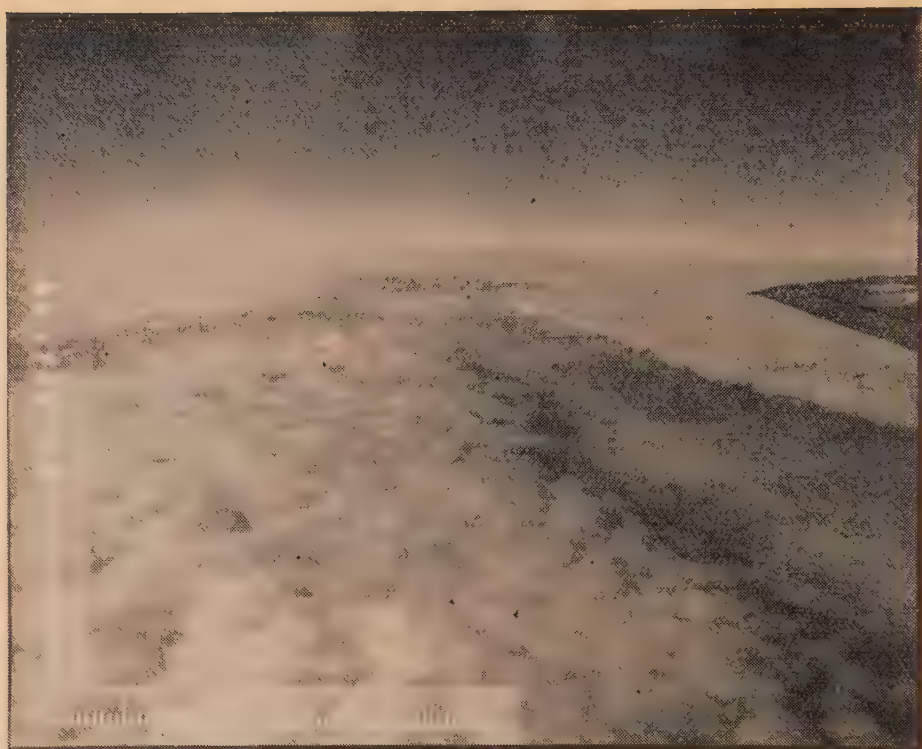


FIG. 21. — Photo of a supercooled stratus cloud which has been seeded with about 1 pound of dry ice fragments per mile of flight (Official photo U.S. Signal Corps Eng. Lab.)



FIG. 22. — Photo of modified stratus cloud seeded at rate of about 1 pound of dry ice per mile of flight. The modified area which appears dark is 15 miles long and 2 miles wide. Note the reflection of the sun produced by the plates in the snow area.

in thick and unstable stratiform cloud systems by certain seeding techniques. Studies are now underway by our group to explore these possibilities.

Some Possibilities and Limitations in Modifying Clouds

There are, of course, definite limitations to the effects which may be expected when supercooled cloud systems are artificially modified with a material such as dry ice.

If any appreciable amounts of precipitation are to result from such activities, the atmospheric conditions must be such that it is likely snow or rain would probably occur from natural processes. The method does, however, present the definite possibility of « triggering off » a storm sooner than it would develop if the natural sequence prevailed.

It must be remembered that there is actually only a relatively small amount of condensed water in a cloud. Thus if all of the moisture in an ordinary stratus cloud having a vertical thickness of a mile were to be precipitated only about 2 mm of water would result. Towering cumulus of the congestus type may contain from 1 to 3 cm of precipitable water, depending on vertical height and liquid water content.

While it is not possible to produce ice crystals in clouds using the dry ice method unless they are *cold* clouds, with temperature colder than 0°C , it is obvious that if a *cool* cloud was modified in the supercooled region, the resulting snow crystals would melt and become rain at a short distance below the freezing level thus providing water droplets which would lead to the formation of rain. If Langmuir's chain reaction theory is correct, this effect would quickly lead to a rapid development of heavy rain in large cumulus clouds.

The tremendous quantities of ice crystals which may be produced with dry ice in cold clouds raises the distinct probability that, if correctly introduced, it should be possible to eliminate dangerous hail storms. With a large concentration of sublimation nuclei present in the region where hail might form, the competition between particles for the available moisture would be so keen that no particles could grow large. It would, indeed, be foolish to think that such results could be obtained immediately. Much careful quantitative research experiments must be conducted to hope for eventual success in this field. In a similar manner, violent thunderstorms may eventually be subdued by preventing the development of any appreciable thickness of supercooled cloud in the cold portion of the cool cumulus congestus cloud.

In fact, any type of storm system which contains appreciable amounts of supercooled clouds may be profoundly modified by intelligent dry ice seeding.

Supercooled ground fogs may also be modified by following a proper seeding procedure. There is a likelihood of « overseeding » with ground fog which not only reduces visibility but also makes the fog much more persistent. Proper amounts used will, however, clear the fog quite rapidly.

There is also a real possibility that the aircraft icing problem may be affected to a considerable degree by modifying supercooled clouds in the vicinity of airports and along heavily travelled air lanes. There is also a distinct possibility that the local area ahead of a flying aircraft may be modified with projectiles carrying and dispensing seeding media, such as dry ice from liquid carbon dioxide. In this way, it may be possible that the icing hazard could be reduced or eliminated as the plane reached the treated area.

One of the immediate results which is proving to be of much value in the physical studies of clouds is the fact that the dry ice seeding process permits the marking of clouds in a unique way. By photographing the subsequent appearance of the modified cloud, many interesting features may be measured in a quantitative manner by special photographic techniques and procedures under development by our Project Cirrus operation.

The science of experimental meteorology is just beginning. The next few years will show rapid advances in this fascinating field. The results will surely aid meteorologists and others in reaching a better understanding and appreciation of weather phenomena.

Table I. — *Comparison of general properties of positive and cross-current storms.*

General Properties	Type of Storm	
	Positive-current	Cross-current
Air temperature	0 to 25° F	25 to 35° F
Precipitation forms	3, 4, 5, 7, 8,	1, 2, 6, 8, 9
General nature of snowfall	Dry	Wetish
Clumping tendency	Rare	Common
Frozen water content	0.05 to 0.2 g/m ³	0.05 to 1 g/m ³
Rate of snowfall	0.1 to 1 cm/hr	0.1 to 3 cm/hr
Density on ground	0.05 to 0.15	0.05 to 0.75
Weight of crystals	4 to 100 micrograms	4 to 1000 microgr.
Falling velocity	50 cm sec ⁻¹	20 to 250 cm sec ⁻¹
Range of falling velocity	Small	Large
Rime deposition	Rare	Common
Cloud structure	Nonconvective	Convective
Air-to-ground current	Positive	Positive and negative
Observed atmospheric electricity, C _p collector	Less than 0.25 μ A	—7.3 to +9.7 μ A
Frequency of electrical activity	Sporadic and short	Nearly continuous
Relationship of electrical activity to specific clouds	Not obvious	Often quite definite
Rain forms	Light	Thunderstorms

Table II. — *General features and relative sizes of frozen precipitation.*

1. Stellar crystals Six rayed, starlike forms occurring as single crystals or in clumps; crystals vary in a wide range of form and size, often comprising cottony flakes sometimes two inches in diameter; size 1/32 inch to 1/2 inch; clumps common.
2. Graupel Snow crystals which are covered with frozen cloud particles forming a rime deposit. This results in pellets of soft snow, which range from hexagonal to rounded kneaded forms; size 1/16 inch to 1/4 inch; rarely clumped.
3. Hexagonal plates Thin, hexagonal, solid or semisolid plates often containing internal structure due to air inclusions. The smallest form of this crystal and the stellar form is known as «diamond dust»; size 1/1000 inch to 3/16 inch; rarely clumped.
4. Hexagonal columns Generally flat-ended, sometimes with one end pointed, hexagonal prisms mostly of transparent ice, many with air inclusions; often a group of columns grow from a common source radiating in several directions. The former type produces 22° and 46° halos around the Sun; size 1/64 inch to 1/8 inch; sometimes clumped.
5. Capped hexagonal columns Similar to hexagonal columns with the exception that both ends terminate as expanded hexagonal plates; a hexagonal plate sometimes is positioned midway between the two plates. Many queer forms occur; size 1/64 inch to 3/16 inch; sometimes clumped.
6. Ice needles Long slender shafts with hexagonal cross section often terminating with sharp irregular points on periphery; needles occasionally grow in masses in random directions; size 1/64 inch to 3/8 inch; often clumped.

- 7. Asymmetrical crystals Angular crystals without a symmetrical outline occurring as simple or compound crystals; sometimes crystalline rays originate from a common center but grow in many directions; size 1/64 inch to 1/8 inch; often clumped.
- 8. Powder snow Dry bits of snow without angular form, often of irregular shape; size 1/64 inch to 1/8 inch; rarely clumped.
- 9. Sleet Pellets of translucent to transparent ice often bearing bumps or other protuberances; they are frozen rain drops but are not always round; size 1/32 inch to 1/4 inch; very rarely clumped.

Table III. — *Extremes in number of sublimation nuclei as a function of air temperature.*

Date	Time	Temp °C	N° of Nuclei	Type	Comments
2 /11 /48	0800	—18	> 50	hex.	Alt. st. clouds
2 /24 /48	0700	—18	20 - 30	x	Supercooled fog in low-lands
12 /25 /48	0730	—14	30 - 40	hex.	Clear
1 /23 /48	0800	—14	10 - 15	hex.	Stratus overcast
1 /22 /48	0815	—10	10 - 20	hex. x	Stratus overcast
2 /13 /48	0745	—10	3 - 5	hax. x	Cirrus haze
1 /25 /48	1200	—7.8	20 - 30	hex. x	« Frosty » air
2 /19 /48	0730	—7.8	1 - 5	x	Cirrus haze
2 /15 /48	0800	—6.6	5 - 10	hex. x	0.8 St. cu.
3 / 1 /48	2200	—6.6	none	—	Cirrus haze
2 /24 /48	2300	—6.1	none	—	Clear, 22° halo
3 /15 /48	0645	—6.1	none	—	Cirrus haze

Table IV. — *Variation in number of ice nuclei under nearly isothermal conditions.*

Date	Time	Temp. °C	N° of Nuclei	Type	Comments
2 /11 /48	1300	— 9.5	20 - 30	hex., x	Alt. cumulus
1 /22 /48	0815	—10	10 - 20	hex., x	Thin cu. hum.
1 /23 /48	0745	— 9	10 - 20	hex.	Cu. humilus
2 /23 /48	0800	— 9.5	5 - 10	hex., x	Scat. fr. cu., hazy
2 /13 /48	0745	—10	3 - 5	hex., x	Cirrus haze
3 /11 /48	0645	— 9.5	2 - 5	hex.	0.8 st. cu. sun pillar
2 /22 /48	2200	— 9.5	1 - 3	hex., x	Clear
3 / 6 /48	2130	— 9	1 - 2	hex., as.	Cirrus haze
3 /13 /48	2200	— 9	1 - 2	x	Thin haze
2 /20 /48	0745	— 9.5	0.1 - 1.0	hex.	Clear

Table V. — Variation in number of sublimation nuclei as a fonction of temperature of free air. — Cold chamber method.

Temp.	Ca $1 \times 10^6/M^3$	Ca $1 \times 10^3/M^3$	$< 1 \times 10^2/M^3$	None
— 30° C	4	5	1	1
— 26°	3	3	1	4
— 22°	5	4	0	1
— 18°	6	4	0	1
— 14°	3	5	0	3
— 10°	5	4	0	3
— 6°	1	2	5	4
— 2°	0	1	3	7
+ 2°	0	3	3	6
+ 6°	2	0	6	3
+ 10°	0	2	3	6
+ 14°	0	1	9	1

REFERENCES

1. SCHAEFER, V.J. — « Properties of Particles of Snow and the Electrical Effects They Produce in Storms ». Trans. Amer. Geophys. Union, 28, 587, (1947).
2. SCHAEFER, V.J. — « Final Report on Icing Research up to July 1, 1945 ». General Electric Research Laboratory. A.T.S.C. Contract W-33-038-AC-9151.
3. BENTLEY, W.A. and HUMPHREYS, W.J. — « Snow Crystals ». McGraw Hill Publishing Company, New York City.
DOBROWOLSKI, A.B. — *Historja Naturalna Lodu*, Warsaw (1923).
HELLMANN, G. — *Schnee Krystalle*, Berlin, (1893).
NAKAYA, U. and SEKIDO, Y. — Hokkaido Imperial University, Sapporo, Japan, Ser. II, V. 1, N° 9 (1935).
4. SCHAEFER, V.J. — « The Production of Ice Crystals in a Cloud of Supercooled Water Droplets ». Science, 104, pp. 457-459, (1946).
SCHAEFER, V.J. — « The Production of Clouds Containing Supercooled Water Droplets or Ice Crystals under Laboratory Conditions » Bull. Amer. Met. Soc., 29, pp. 175-182 (1948).
5. SCHAEFER, V.J. — « Method of Making Snowflake Replicas ». Science, 93, pp. 239-240 (1941).
SCHAEFER, V.J. — « Use of Snowflake Replicas for Studying Winter Storms ». Nature, 149, p. 81 (1942).
6. SCHAEFER, V.J. — « The Detection of Ice Nuclei in the Free Atmosphere ». To be published by American Meteorological Society.
7. FINDEISEN, W. — « Colloidal Meteorological Processes in the Formation of Atmospheric Precipitation ». Met. Zeitschr., 55, pp. 121-133 (1938).
8. VONNEGUT, B. — « The Nucleation of Ice Formation by Silver Iodide ». Jour. Appl. Phys. 18, pp. 593-595 (1947).
9. LANGMUIR, I. — « The Production of Rain by a Chain Reaction in Cumulus Clouds at Temperatures above Freezing ». Jour. of Meteorology (Aug. 1948).
10. NAKAYA, U. et al — « Further Experiments on the Artificial Production of Snow Crystals ». Jour. Fac. of Science, Hokkaido Imperial University, Ser. II, V. II, N° 1.
11. VONNEGUT, B. — « The Production of Ice Crystals by the Adiabatic Expansion of Gas ». Science, 107, (1948).
12. « First Quarterly Progress Report, Meteorological Research ». Obtainable from Office of Technical Services, Department of Commerce, Washington 25, D.C.
SCHAEFER, V.J. — « Methods of Dissipating Supercooled Clouds in the Natural Atmosphere ». Jour. Inst. of Navigation, Vol. 1, N° 7, 8, pp. 172-174 (1947).
SCHAEFER, V.J. — « Experiments with Cloudy Skies ». Weatherwise, Vol. I, pp. 3-5 (1948).
SCHAEFER, V.J. — « The Natural and Artificial Formation of Snow in the Atmosphere ». To be published in Trans. Amer. Geophys. Union.

IV. — Rapports relatifs au :
GRAIN DE GLACIER



CONTRIBUTION A L'ETUDE DU GRAIN DE GLACIER

Note communiquée au Congrès d'Oslo de la Commission internationale
de la neige et des glaciers
(19-28 août 1948)

par
le Prof. **André RENAUD** (Suisse)

1. Introduction

La glace des glaciers résulte de la lente métamorphose de la neige qui s'est accumulée dans les névés-réservoirs. La structure de cette glace est bien connue et je me bornerai à rappeler ici qu'elle est formée de « grains » de glace dont chacun est un cristal du système hexagonal, mais de surface très irrégulière. Enchassés les uns dans les autres, sans orientation privilégiée (tout au moins dans les glaciers alpins) de leur axe principal, ils forment à une température inférieure à leur point de fusion une masse si compacte qu'un choc produit une cassure conchoïdale qui ne tient pas compte des surfaces de contact.

La fusion de la glace de glacier a été minutieusement décrite par maints auteurs depuis le suisse Hugi (1843) et je me limiterai à en retracer ici très sommairement le processus. Un bloc de glace exposé aux rayons du soleil ou à ceux d'une autre source de radiations calorifiques ne fond pas d'une manière homogène. La fonte se produit d'abord à la surface des grains, et les cristaux ainsi dissociés ressemblent quelque peu aux pièces enchevêtrées d'un jeu de « puzzle ». Plus tard, les grains séparés plus complètement les uns des autres montrent de nouvelles phases de la fusion qui font apparaître les stries de Forel et les figures de Tyndall. Les stries de Forel séparent, à la surface du cristal, les lamelles de glace qui sont disposées perpendiculairement à l'axe principal, et les figures de Tyndall sont le plus souvent des lentilles de fusion à l'intérieur des lamelles, contenant souvent aussi un peu d'air.

Ainsi, lorsque la température du glacier s'approche du point de fusion, la masse perd sa compacité et devient un agrégat de cristaux de glace séparés par un film liquide. Si, par suite de la fusion, cette pellicule n'est plus maintenue en équilibre à la surface des grains, elle s'évacue par les fissures capillaires observées déjà par les anciens naturalistes, fissures qui ne sont en réalité que des canaux inter-granulaires résultant d'une fusion avancée.

Les faits d'observation rappelés ci-dessus sont classiques et laissent présumer l'existence d'une constitution hétérogène du grain de glacier au sujet de laquelle diverses hypothèses physico-chimiques ont été émises (voir 3), et m'ont incité à une étude expérimentale dont cette communication donne un aperçu résumé.

2. Expériences sur la structure physico-chimique du grain de glacier

De 1935 à 1947, j'ai prélevé environ 50 échantillons de glace sur les glaciers suisses de Gorner, Ried (Valais), Fee, Rhône. Saleinaz et Aletsch. Ils proviennent généralement de la langue glaciaire, mais la Station scientifique du Jungfrau-Joch m'a permis également d'étudier la glace de la région supérieure du glacier d'Aletsch. Chaque échantillon a fait l'objet d'une fusion fractionnée, soit au soleil, soit au rayonnement de l'arc électrique, et les diverses tranches successives d'eau recueillies ont été immédiatement analysées pour en déterminer le degré de pureté. La faible teneur d'impuretés que ces eaux ont révélée lors des essais a écarté toute possibilité pratique d'analyse gravimétrique et j'ai adopté la méthode de la détermination de leur conductivité électrique comme entrant seule en considération dans un travail qui ne pouvait pas toujours s'effectuer dans un laboratoire. Cette méthode est trop connue pour être décrite ici, mais il convient cependant de préciser que, dûment adaptée à une recherche qui exigeait un degré de précision élevé, elle a donné d'excellents résultats. Ceux-ci sont consignés plus loin (voir 3) et se rapportent à la conductivité de l'eau en Mhos/cm. Etant donné la très faible concentration des sels qui causent cette conductance, on peut admettre que la conductivité est proportionnelle à leur concentration. D'autre part cette dernière n'a pas encore été calculée, car mes recherches n'ont pas encore porté sur la nature des ions contenus dans la glace de glacier. Il est cependant des plus probables qu'ils sont apportés par les précipitations atmosphériques dans lesquelles on reconnaît, selon divers auteurs : l'azote ammoniacal, nitreux et nitrique, des chlorures, sulfates, le sodium et des gaz dissous. D'après mes propres observations, la concentration de ces sels est du même ordre

dans la neige fraîche, la pluie et la glace de glacier; et l'acide carbonique ne se rencontre pas dans la glace.

J'ai généralement fondu 300 g de glace dont j'ai recueilli séparément les tranches d'eau successives de 30 g chacune. La première provient indiscutablement de la surface des grains, et la dernière de leur partie centrale. Bien que les masses de ces tranches soient égales, elles correspondent à des épaisseurs bien différentes. La première ne représente que les 0,035 du diamètre du grain, soit environ le trentième, tandis que la dernière en constitue les 0,465 soit à peu près de la moitié. La première tranche fondue correspond donc à une pellicule relativement mince, mais sa séparation complète et rapide des couches sous-jacentes s'est révélée très difficile. Il est donc probable que, dans nombre de cas, une certaine diffusion des eaux de fusion peut se produire et la conductibilité de la tranche recueillie peut ainsi différer quelque peu de celle de la tranche fondue. Enfin, l'évacuation de l'eau contenue dans les lentilles de Tyndall altère d'autre part les résultats. Cependant, sans être négligeables, ces deux facteurs ont peu d'importance et n'influencent d'aucune manière les résultats généraux et les conclusions théoriques.

3. Principaux résultats

a) Le grain de glacier est toujours constitué d'un cristal de glace pure entouré d'une pellicule saline. Quelle que soit la provenance de la glace, ce fait ne comporte aucune exception.

b) La salinité de la pellicule est plus grande dans la glace du névé que dans celle du glacier proprement dit.

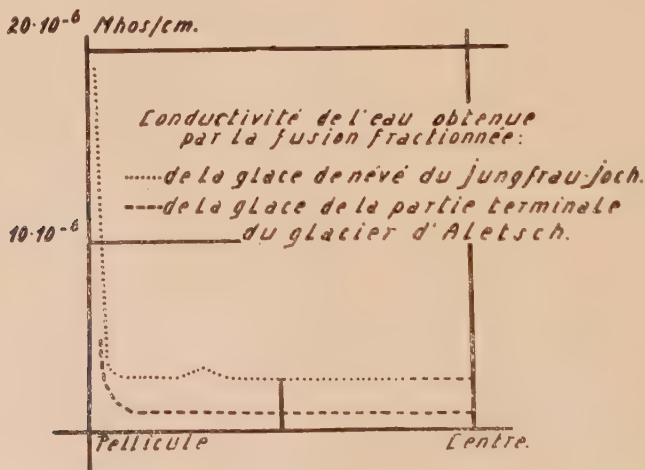
Exemple : Au Jungfrau-Joch (altitude 3450 m.) j'ai étudié la glace froide (à une température moyenne de -6° à -4° formée de grains de petites dimensions (maximum 1,5 à 2 cm) dont la conductivité de l'eau issue de la pellicule est en moyenne de $30 \cdot 10^{-6}$ Mhos/cm. (max. $120 \cdot 10^{-6}$ Mhos/cm)

Dans la région inférieure du glacier d'Aletsch, à l'altitude de 1800 m environ, j'ai étudié une glace constituée par des grains de dimensions beaucoup plus considérables et à une température plus élevée qui atteint le point de fusion durant la plus grande partie de l'année. La conductibilité de l'eau de fusion de la pellicule saline de cette glace n'est plus, en moyenne que de $4 \cdot 10^{-6}$ Mhos/cm (max. $12 \cdot 10^{-6}$ Mhos/cm.)

c) La pureté de la glace qui constitue le noyau du cristal est plus grande dans la glace de la région terminale que dans celle du névé.

Dans l'exemple du glacier d'Aletsch (litt. b), j'ai trouvé une conductivité moyenne de l'eau issue de cette glace égale à $3 \cdot 10^{-6}$ Mhos/cm au J'Joch (minimum $1,5 \cdot 10^{-6}$ Mhos/cm), contre $1 \cdot 10^{-6}$ Mhos/cm (min. $0,85 \cdot 10^{-6}$ Mhos/cm pour la glace du bas du glacier).

Je relève en passant que l'eau fournie par le centre des grains de glacier a souvent une pureté comparable à celle des meilleures eaux distillées qu'on peut préparer et conserver en laboratoire.



Graphique 1.

d) Entre ces valeurs extrêmes, la conductivité des tranches intermédiaires varie selon une fonction dont le graphique 1 donne une idée.

e) Si les résultats varient beaucoup selon les échantillons, il n'y a par contre aucune différence systématique entre la glace des bandes blanches et celle des bandes bleues. Il est erroné de prétendre, comme on l'a fait souvent, que l'une de ces glaces est plus pure que l'autre.

4. Conclusions

Les résultats ci-dessus apportent la preuve de l'hétérogénéité physico-chimique du grain de glacier. Ils constituent, à ma connaissance tout au moins, la première vérification expérimentale de la théorie de Buchanan (1) selon laquelle la glace naturelle est l'association d'un cristal de glace pure et d'une saumure saline adhérente. On sait que cet auteur avait justifié cette hypothèse par une étude très poussée de la température de congélation d'une solution saline à diverses concentrations et par l'étude du coefficient de dilatation thermique de la glace obtenue dans ces conditions et à différentes températures. Les résultats de ce travail confirment également une hypothèse analogue formulée par Quincke (2). Ils permettent d'écarter la théorie du ciment amorphe appliquée à la glace (3).

Un examen plus attentif des résultats donnés plus haut (voir 3) montre qu'au cours de leur croissance et lorsqu'ils atteignent les régions chaudes du glacier où s'effectue leur fusion, les grains de glacier éliminent une grande partie de leur saumure, et cela plus rapidement dans la pellicule que dans le centre des grains.

Ces recherches n'ont permis d'étudier jusqu'ici que des glaces superficielles ou presque. Quelques grottes m'ont livré de la glace plus profonde. J'ai l'intention de poursuivre mes mesures sur de la glace prise à de plus grandes profondeurs, si possible jusqu'au lit glaciaire. Dès lors, il sera possible d'être mieux renseigné sur l'évolution physico-chimique des grains de glacier au cours de leur voyage vers les régions basses et de tirer des conclusions plus générales.

5. Quelques conséquences théoriques

Bien que mes recherches ne soient pas terminées et n'autorisent pas encore de formuler de nouvelles hypothèses sur le mécanisme du glacier, il me semble cependant utile d'exposer quelques vues personnelles sur les questions posées par le phénomène glaciaire.

Il me semble tout d'abord que le problème de la croissance des grains ne pourra pas être résolu sans que des observations sur la température du glacier dans ses différentes parties aient été enfin réalisées dans nos Alpes. D'autre part, des recherches de laboratoire sur la formation de la glace de glacier seront nécessaires. Je présume que cela permettra de montrer que la croissance des cristaux dépend essentiellement de leur température, spécialement au voisinage du point de fusion, car l'élimination de la pellicule saline entre deux cristaux voisins doit faciliter leur coalescence. L'orientation des cristaux et la pression jouent peut-être un rôle dans ce processus de croissance qui a le caractère d'une soudure autogène, sans modification dans l'équilibre thermique.

Enfin, l'existence probable d'un film liquide enrobant les cristaux dans la plus grande partie de la masse du glacier tempéré doit jouer dans le déplacement des grains le rôle d'un lubrifiant et permettre de trouver enfin une explication de la « semi-fluidité » du glacier entrevue déjà par Agassiz : « Le glacier... est une agglomération de fragments de glace rigides, juxtaposées et rendus mobiles par l'introduction d'un corps étranger dans leurs interstices ». Le fait que la température joue un rôle essentiel avait été déjà entrevu par cet auteur qui ajoutait que « Le mouvement du glacier... est dans la plus étroite dépendance du climat, puisque sa progression varie suivant les saisons ». (Nouvelles études et expériences sur les glaciers actuels 1847). Toute hypothèse nouvelle sur la « plasticité » ou la « viscosité » du glacier devra tenir compte du rôle du film liquide produit par la fusion de la pellicule saline que possèdent les grains.

Bibliographie :

- (1) J. Y. BUCHMANN, M.R.I. Ice and its natural history. 1908.
- (1) G. QUINCKE. Ueber Eisbildung und Gletscherkomm. 1905
- (1) HAWKES. Some Notes on the Structure and Flow of Ice. 1930.

2) THE GROWTH OF THE GLACIER CRYSTAL

by

GERALD SELIGMAN

Introduction

About fifteen or twenty years ago I and others recorded the evolution of the snowflake into a crystal of firm snow. In 1938 the Jungfrauoch Glaciological Research Party under my leadership continued this investigation as far as the transition of firm snow into glacier ice. At this stage the crystals, more properly called crystallites, had diameters of about 0.5 to 1.75 mm. This already indicates considerable growth, for the new snow grains, once they have lost their dendritic structure, seldom exceed 0.3 mm. in diameter. In the tongues of glaciers the crystals are of centimetre rather than millimetre size.

It was desired to investigate these later stages, thus completing the whole story of crystal growth in glaciers. The ultimate aim is to establish the mechanism of growth. In the case of snow this is tolerably well known. So far as the later stages of ice are concerned little is known for certain. Much work has still to be done in the field and in the laboratory.

The lecturer then showed a number of slides illustrating this transition.

As a preliminary to the last stages of this investigation, the growth of the large crystals in the glacier tongue and in dead ice, it seemed necessary to have some definite information about the size of these crystals. Also to find out where the growth is greatest and to discover what conditions might affect growth.

This paper deals with these points and presents certain facts which will be useful in approaching the ultimate theme, the solution of which would not only be of value glaciologically, but might also throw light on geological problems, not only those of rock metamorphism, but also those of the Pleistocene. It would be particularly important if this research enabled us to tell the age of dead ice formations. It might also prove of metallurgical interest.

METHODS

For the purpose of the present research a very quick method of recording crystal sizes had to be devised. Friends at the Weissfluhjoch, Davos, suggested taking rubbings — a method originally used by Deeley and by Forel.

Crystal boundaries are more sensitive to weathering than the rest of the ice and become depressed below the general surface. If a piece of soft printing paper is laid over the ice and rubbed with a soft pencil the boundaries remain white, as will be seen presently.

In the Alps the crystal boundaries of surfaces which have been exposed for some time are usually visible. If a block of ice is freshly cut it is necessary to iron it with a meta iron before they show up. A useful alternative, recommended to me by Dr. Mercanton is to place a flat metal plate on the ice. This absorbs radiation and, when removed, leaves a flat surface ready for rubbing. Professor Ahlmann has devised another method and it is safe to say that the technique of reproducing crystal boundaries will undergo further improvements soon.

Rubbings of course only give a two-dimensional picture. This is not a very serious error in actively flowing ice as the crystals usually measure about the same on their three main axes. In dead ice this is not the case, but in any case the crystals are generally too large for rubbings.

A more serious error is that sections cut through an assemblage of crystals will not pass through the largest diameters of every crystal. Results therefore tend to be low. It may, however, be assumed that over the considerable number of crystals measured (some 16,000 in the case of the present research) the error is fairly constant.

For the more detailed survey which must follow this research some more accurate method may have to be devised or some correction applied (see for instance the work of W. A. Johnson, Estimation of spatial grain size, *Metal Progress*, January 1946, p. 87-92).

Minute crystals cannot be detected by the rubbing method. I found that a floor stain, Darkalene, showed these up very well, but it seems that they are seldom to be found in quantities in glacier tongues.

THE CRYSTAL COUNT

For measuring the crystals on the rubbings, after prolonged tests the following method was found to give the most satisfactory results.

A series of standard circles of increasing size were drawn on tracing cloth. This was placed over the rubbing and the number of two-dimensional crystal surfaces corresponding to their equivalent circles was counted. The sum of the areas of each size of equivalent circle represented in the rubbing was then expressed as a percentage of the whole area of the sample. A histogram is a convenient method of showing the proportions of crystals of each size present. The root mean square diameter of the standard circles gives the average crystal diameter of the whole sample. In this paper whenever a mean diameter of an assemblage of crystals is given it refers to the root mean square diameter.

The standard diameters were arranged logarithmically so as to give a constant ratio between each size and the next. They were : 0.25, 0.4, 0.6, 1.0, 1.6, 2.5, 4.0, 6.3, and 10.0 cm. They are rounded off to not more than two significant figures, the deviation being within the error of experiment, since each crystal can only be approximately matched with its equivalent circle.

Crystal size v. length of travel

The next slide shows the mean crystal sizes at the ends of several Swiss Glaciers plotted against the length of the inclined travel of the glacier. It will be seen that they increase proportionately with the glacier length.

Crystal size v. angle of slope

The next slide shows the same crystal sizes plotted against the mean angle of slope of the same glaciers. Here it will be seen that they decrease in inverse proportion to the gradient.

Plan of the Eiger Glacier Snout

The next slide shows the mean crystal sizes in the end of the Eiger Glacier. Those in the centre of the main stream average only 1.43 cm. It should be noted that they increase towards the sides where the ice flows slower.

One authority has tried to show that this is due to greater stresses being set up by friction at the glacier banks. I hope that I shall be able to show that this is not a necessary hypothesis.

DEAD ICE

This increase in crystal size with decrease of speed reaches its maximum in dead ice. The largest measured was $18 \times 8 \times 5$ cm. Crystals like those in the next slide were not uncommon.

In the Alps the largest masses of dead ice are to be found beyond the ends of the glaciers where the ice has been as it were left high and dry by the present recession. In some glaciers, however, the shrinking ice leaves dead ice deposits at its margins, halted there through the friction of the bank.

The next slide shows a case of this sort in the Findeln Glacier with the active ice making a pronounced break and sweeping past the dead ice.

SUB-NORMAL CRYSTALS

So far we have seen a gradual increase in crystal size, but there seems to be an exception. In a tunnel dug into the ice near the end of the Upper Grindelwald Glacier, we found crystals of subnormal size the smallest assemblage measuring

0.47 cm. It is important to note that this tunnel lies below a great ice fall — a point to which will refer later.

I have never observed crystals of this small size so near the end of the glacier. However, search through old writings brought to light a paper by Deeley (Deeley R. M. The structure of glacier ice. *Geological Mag.* New Series, Decade V, Vol. 2, 1895, p. 157) in which he records subnormal crystals in tunnels in the Lower Grindelwald and the Morteratsch Glaciers. I subsequently found small crystals in a tunnel in the Lower Grindelwald Glacier but they were not so small as those of the Upper Grindelwald Glacier, the smallest assemblage averaging 0.77 cm. Various explanations suggest themselves :

1. This phenomenon has only been observed in artificially excavated tunnels and may also be artificial in origin.

2. It cannot be regarded as wholly unlikely that down in the interior of the glacier crystal growth is much slower than at the surface. Also that these subnormal crystals are the undeveloped crystals of the basal ice opened up to inspection by the excavation. This is believed to be unlikely, but the point is referred to below. Against this the following is to be noted. In the chamber of the tunnel some 24 m from the mouth, there was a stratum of ice with an average crystal size of about 1.6 cm that is to say about normal for its position in the glacier. Its width was about or 2 m. Its thickness could not be ascertained as its base lay beneath the floor. One might well ask what this mass of large crystals was doing here surrounded by small ones, and why they had grown so much larger than the others of the small size if the latter was due to inhibition of growth.

3. We must not overlook the possibility of actual comminution by pressure or shearing, but much doubt is cast in this hypothesis by crystallographers and metallurgists.

4. Again these small crystals might stem from snow falling into crevasses and becoming sealed up in the glacier. They would thus be younger than the rest of the glacier crystals.

5. A more likely explanation seems to be that the forces exerted by the massive ice fall set up stresses which cause recrystallisation. There is ample evidence for this in the behaviour of worked metals in which new crystal nuclei develop. Tammann, discussing the behaviour of camphor, writes, « ...Bei stärkere Deformation bilden sich viele Rekristallisationszentren. » Later in the same paper he hints at a similar behaviour of ice immediately after being rolled in the laboratory.

There is another analogy with worked metals. These small crystals under suitable conditions, grow very fast. In the Upper Grindelwald Glacier the mean crystal size in the snout is 1.79 cm, betokening very rapid growth between the tunnel and there, the distance being a matter of only a few score metres. Nowhere else was anything approaching this speed of growth observed.

Against the concept of recrystallisation the views of Bader must be cited. Bader subjected a block of ice to shear and found no development of new nuclei. But as against 0° C. in the living glacier his experiments were conducted at — 5° C.

Clearly the phenomenon needs more detailed investigation than was possible on this occasion.

Summing up, the following facts have come to light :—

1. The increase in crystal size observed at or near the surface from bergschrund to snout of an Alpine glacier is a gradual one.

2. In active flowing ice the crystals are smallest on the lines of fastest flow, that is in the centre of the stream. They become larger towards the margins, the increase being gradual.

3. The longer the glacier the larger the crystals.

4. The steeper the slope of the glacier the smaller the crystals.

5. The largest crystals in the actively flowing ice in the centre of the glacier streams in the Alps do not exceed 3 cm. in mean diameter.

6. Crystals may grow to very large size in dead ice even under conditions where no movement can take place.

7. Under certain conditions abnormally small crystals are found in glacier tongues.

8. The rate of crystal growth increases with the temperature. (See below).

These facts lead to the conclusion that, other things being equal, crystal growth in glacier tongues is a function of time.

This confirms the belief expressed by some of the older glaciologists, who, however, offered too little observational proof for acceptance. Although the size of the crystals increases with age, this does not explain the mechanism of growth. But it will be helpful for the future investigations, to which the present research is intended as an introduction, if we examine some of the factors which may influence the mechanism.

It should be pointed out that the present research was conducted at or near the glacier surface. Nowhere were depths lower than 6 m. explored. Little is known of conditions deep down in the ends of glaciers. Excavations a few metres down showed no perceptible variation in crystal size or habit. In the firm area excavations carried out in 1938 showed an increase in grain size down to 19 1/2 m. and thereafter no very marked change. All that can be said, therefore, is that according to the now accepted Finsterwalder flow theory, the large crystals at or near the surface of the snout were once deep down in the glacier. They may have grown to large size as they approached the surface, or they may have grown big in the glacier depths.

THE INFLUENCE OF TEMPERATURE

From such not very copious details as are available the crystals in polar regions seem to be smaller than those in the Alps. This seems to indicate that radiation is of importance in growth, a belief held by many.

Crystal Sizes at the Jungfrauoch

A striking confirmation of this was found at the Jungfrauoch. There is a very large mass of ice forming part of the Sphinx ice apron, which lies immovable in a kind of pocket. A series of large halls and long corridors has been excavated in this. They are lit by electric lights on brackets attached to the ice walls. The normal crystal size throughout the excavations was of the order of 0.29 cm. The air temperature was -3 to -4°C . but near the lights measurements showed a mean size of 1.07 cm. The size tailed off as the distance from the lights increased. The bulk of the ice being at -4°C . crystal growth tends to be slow. On the neighbourhood of the lamp, however, the ice is heated up probably to 0°C . This rise in temperature accelerates crystal growth in the ice as it does in all other substances.

The process can be compared with annealing in stressed metals which is accompanied by crystal growth. There are no doubt strains upon the ice crystals in the roof of the excavation which give rise to crystal growth when these are annealed by the heat of the lamp. Further confirmation of the effect of heat is found in the fact that where melt water played over the ice the crystals were larger than normal.

There was no sign of melt water either close to the lights or by percolation from the surface 10 or 20 mm. above.

Crystal growth in solid ice can therefore take place without visible melting as it certainly does in snow, a point that is confirmed in some polar glaciers.

THE INFLUENCE OF THE FREEZING OF MELT WATER

This thaw-freeze process is nevertheless a tempting one to adopt in order in part to account for crystal growth. It is unquestionably operative in the firm region but in glacier tongues night frosts are unlikely to penetrate any distance below the surface.

The winter cold wave is unlikely to penetrate more than a few metres down owing to an insulating mantle of snow and the surface layers exposed to it will melt away during the summer. Therefore none of the large crystals which are visible for instance in late summer, can have been subjected to the winter cold wave.

It seems, therefore, that melt water in cold regions (as for instance at the Jungfrauoch) can only influence crystal growth by reason of its temperature bringing ice to the melting point and thus stimulating crystal growth, just as does warm air temperature.

THE INFLUENCE OF GLACIER MOVEMENT

Tammann and others believed that movement was a necessary condition of crystal growth. In a paper of which I was part author it was suggested that crystals, so orientated in the glacier that they could yield to shear stresses caused by its movement, would possess less free energy than others, and would grow at their expense by the transfer of molecules across the crystal boundaries.

If this view is correct, it would seem that its effect would decrease towards the snout where it is believed the crystals become preferentially orientated. No such decrease in speed of growth was observed. Moreover, it must not be forgotten that crystals in stationary dead ice grow to far larger sizes than those in moving, active ice.

It seems therefore that movement is not an essential condition of growth.

CONCLUSION

From all these facts it seems to follow that the principal influence on ice crystal growth is warmth and that the longer the crystals are subjected to warmth the larger they grow.

FUTURE RESEARCH

Stated briefly this should be :

1. Detailed crystallographic examination of a single smooth-flowing glacier, and below a single ice fall, and in dead ice.
2. Temperature distribution in a glacier tongue and in dead ice.
3. The recording of crystal sizes in a single mass of dead ice over a term of years, both under controlled laboratory conditions and in the field.
4. Meteorological observations at glacier tongues.
5. Laboratory experiments of shear etc, with real glacier and other ice.
6. The effect of dissolved salts on inhibiting crystal growth.

ACKNOWLEDGEMENTS

I am indebted to many friends who have helped me with advice on various occasions, in particular to my former colleague Dr. M. F. Perutz; also to Mr. H. S. Peiser and Mr. G. Hattersley-Smith.

3) GLACIER ICE CRYSTAL MEASUREMENTS AT KEBNEKAJSE, SWEDEN

by

H. WILSON AHLMANN (Stockholm) and E. G. DROESSLER (Washington D.C.)

After reading Mr. G. Seligman's preliminary report on this subject (1) we initiated a programme for recording and measuring the ice crystals on some of the small valley glaciers in the Kebnekajse *massif* in Swedish Lapland from 18 July to 12 August 1948. This is the region in which Geografiska Institutet, Stockholm, has been carrying out systematic glaciological studies since 1946.

It was very difficult or impossible to use Mr. Seligman's method, chiefly because of the humid weather. The ice surfaces were so wet that the paper became soaked and would not take the pencil marks. Using the limited facilities available at the camp, we experimented with another method. We used a roll of ordinary absorbent kitchen paper towelling and a mixture consisting of clay and a brown powder stain insoluble in water. After this mixture has been rubbed over the selected ice surface a length of the paper is carefully placed over the treated area and gently pressed with the finger tips into close contact with it. When the outlines of the crystals can be seen through the paper, usually after the lapse of two or three minutes, the paper is gently removed from the ice, rolled up with another and carried back to the camp for drying and analysis.

Clear ice relatively free from air bubbles should be sought. Satisfactory records can be made on either vertical or horizontal surfaces, but the best results are obtained on horizontal surfaces where the melt water can drain down into the interstices between the crystals. Considerable difficulties were encountered in making a good impression of the lower part of vertical cross-sections because of the abundance of melt water on the crystals. In general it can be said that the drier the ice surface the better the result.

If the ice surface is suitably weathered the best results may be expected if the clay mixture is spread in a thin, even layer over the crystals. The crystal boundaries show up as clean white lines and the crystal surfaces as rich, evenly coloured areas. (Fig. 6,). If the ice is only slightly weathered the clay mixture must be rubbed vigorously over the ice to force the stain into the boundary lines, the impression of which will then be shown as dark lines and the crystal surfaces as lightly coloured areas.

When a suitable ice surface is not available the ice is to be treated. To make the crystal boundaries distinct enough for an impression we used a small blow-lamp to heat the ice surface. Within a very short time the boundaries appeared and had sufficient depth to be satisfactorily recorded.

The form and size of the crystals can be obtained directly from the impressions with a rule. A better method is to make a master tracing of a single impression on transparent paper and to take the measurements from this, or else by comparing and checking two or more impressions *of the same ice exposure*.

An alternative to the clay and stain method which immediately suggests itself is to use photography instead of paper for recording and measuring the crystal sizes.

A continued series of experiments should be carried out with many different kinds of paper, stain and clay in order to arrive at the optimum combination of all three. Such investigations are planned for the glaciological research group at Kebnekajse.

Investigations were first made in a cave in the stagnant marginal part of Isfallsglaciären. Impressions were taken from 6 blocks (A-F, Table I) which had fallen from the roof of the cave.

Table I gives a summary of the crystal size values. On the impressions we measured the largest and smallest axes of all the crystals and computed their areas. The total number of crystals is shown in Col. 2. The axes and the areas of the largest and the smallest crystal in each sample are shown in Cols. 3, 4, 6 and 7.

(*) « The recording of glacier crystal sizes, 1948. » (Privately circulated.)

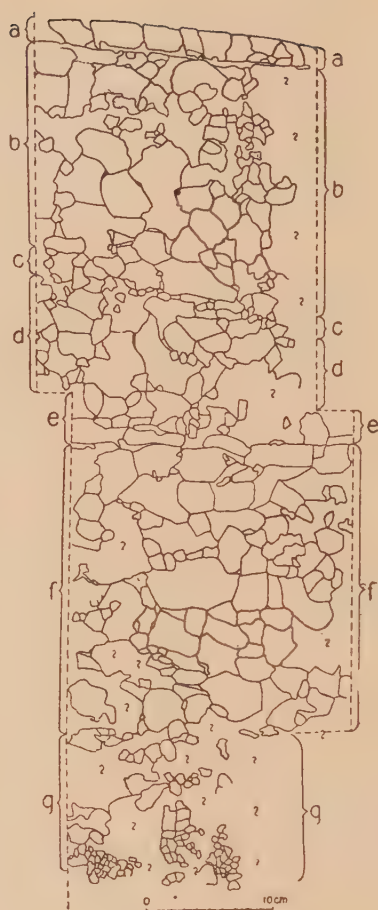
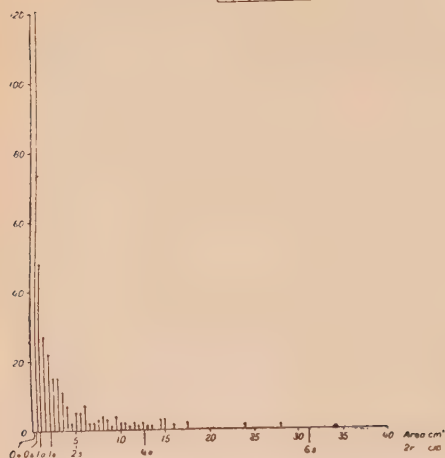
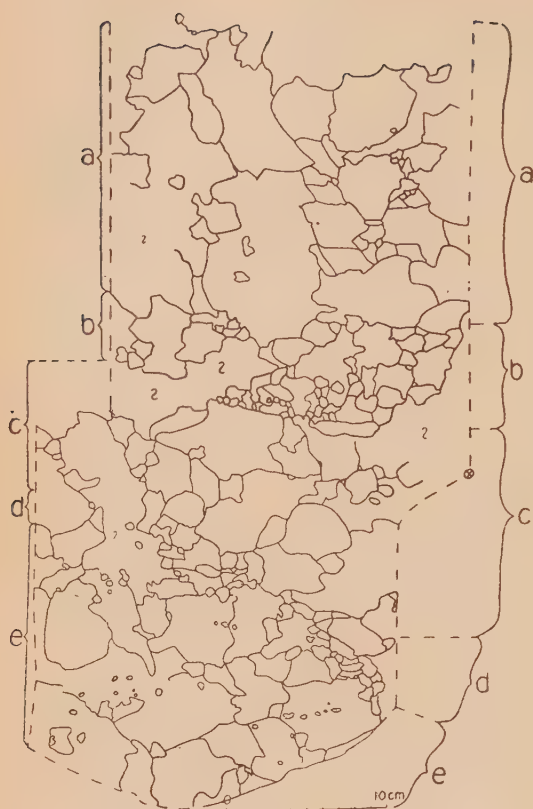


Fig. 2. Top left: crystal pattern of section of ice in the floor of a cave in the margin of Isfallsglaciären

Fig. 4. Top right: crystal pattern in the margin of the tunnel of Storglaciären (Table III)

Fig. 5. Bottom left: Frequency diagram of crystal sizes in Storglaciären

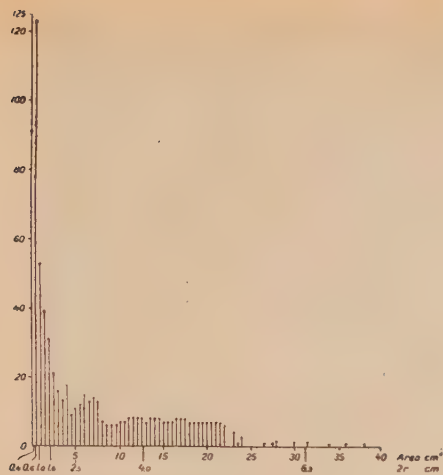


Fig. 1. Frequency diagram of crystal sizes in Table I

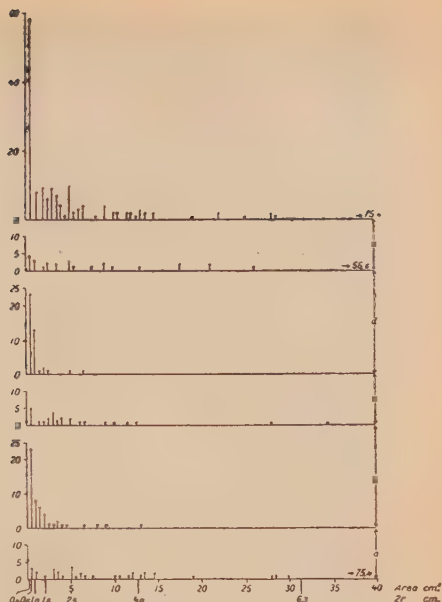


Fig. 3. Frequency diagram of crystal sizes in Table II

The average area of the crystals is given in Col. 9. Because of Seligman's use of diameters as measures of the crystal size we have also calculated these, corresponding to their respective areas considered as circles. (Cols. 5, 8 and 10)

Table I

(1) Block	(2) Total Number of Crystals	Largest Crystal			Smallest Crystal			Average Crystal Size	
		(3) Axes, cm.	(4) Area, cm ²	(5) Dia- meter, cm	(6) Axes, cm.	(7) Area, cm ²	(8) Dia- meter, cm	(9) Area, cm ²	(10) Dia- meter, cm
A	27	7.1 × 3.1	22.0	5.3	0.5 × 0.2	0.1	0.76	4.6	2.4
B	88	7.3 × 3.6	26.4	5.8	0.3 × 0.3	0.1	0.76	2.3	1.7
C	54	11.2 × 4.5	50.4	8.0	0.3 × 0.2	0.1	0.76	5.7	2.7
D	82	8.5 × 4.5	38.7	7.0	0.4 × 0.2	0.1	0.76	3.6	2.1
E	60	7.9 × 4.0	31.6	6.4	0.5 × 0.1	0.1	0.76	5.0	2.5
F	225	9.0 × 4.0	36.0	6.8	0.2 × 0.1	0.02	0.16	2.8	1.9
Total	536							3.5	2.1

In order to give a more detailed description of the crystallographic structure of the ice examined, frequency diagrams of crystal sizes have been calculated, using the intervals : 0.1-0.2, 0.3-0.7, 0.8-0.2 cm², etc. We have also shown as abscissae the diameters selected by Seligman, namely 0.25, 0.4, 0.6, 1.0, 1.6, 2.5, 4.0, 6.3 and 10 cm. Fig. I summarizes all crystals measured on the six blocks.

A vertical cross-section was made through the second layer of ice from the floor of the cave. Six impressions of the cross-section were made and used in arriving at the crystal pattern shown in Fig. 2. It can be divided into five zones. Table II summarizes the values found and Fig. 3 shows the frequency diagram.



FIG. 6. Impressions of crystals on Kebnepakteglaciaren. A 5 cm. long line in the bottom left corner shows the scale.

Table II

(1) Zone	(2) Total Number of Crystals	Largest Crystal			Smallest Crystal			Average Crystal size	
		(3) Axes, cm	(4) Area, cm ²	(5) Dia- meter, cm	(6) Axes, cm	(7) Area, cm ²	(8) Dia- meter, cm	(9) Area, cm ²	(10) Dia- meter, cm
a	40	11.1 × 6.8	75.4	9.8	0.3 × 0.3	0.1	0.3	9.9	3.5
b	66	5.1 × 2.4	12.2	7.0	0.3 × 0.3	0.1	0.3	1.4	1.3
c	27	8.0 × 4.3	34.4	6.6	0.5 × 0.3	0.2	0.4	6.0	2.8
d	52	4.2 × 1.6	6.7	2.9	0.2 × 0.2	0.04	0.2	0.9	1.1
e	27	11.3 × 5.0	56.5	8.5	1.1 × 0.5	0.6	0.3	0.3	3.4
Total	212							4.5	2.4

A cross-section through an ice layer in a transverse crevasse in the margin of the tongue of Storglaciären is shown in Fig. 4. The glacier is 3 km long and the ice was nearly stagnant at the location investigated. Table III summarizes the crystal size values and Fig. 5 shows the frequency diagram.

Table III

(1) Zone	(2) Total Number of Crystals	Largest Crystal			Smallest Crystal			Average Crystal Size	
		(3) Axes, cm	(4) Area, cm ²	(5) Dia- meter cm	(6) Axes, cm	(7) Area, cm ²	(8) Dia- meter, cm	(9) Area, cm ²	(10) Dia- meter, cm
a	15	4.0 × 2.1	8.4	3.3	0.6 × 0.3	0.2	0.5	2.4	1.8
b	102	8.0 × 3.0	24.0	5.5	0.5 × 0.4	0.2	0.5	3.0	1.9
c	33	1.7 × 1.1	1.9	1.5	0.3 × 0.3	0.1	0.3	0.6	0.9
d	47	5.3 × 3.0	15.9	4.5	0.5 × 0.5	0.3	0.6	3.2	2.0
e	18	3.0 × 1.9	5.7	2.7	0.6 × 0.3	0.2	0.5	1.4	1.3
f	97	5.0 × 3.0	15.0	4.7	0.5 × 0.2	0.1	0.4	3.8	2.2
g	86	1.7 × 1.1	1.9	1.5	0.2 × 0.2	0.04	0.2	0.5	0.7
Total	398							2.3	1.7

A superficial measurement of a sample of clear ice from the central part of the tongue of Storglaciären (2.5 km from its source, and in close proximity to Stake No. 24) showed the crystal size to be approximately 3-3.5 cm². The movement of the glacier at the point is about 1.5 cm a day.

A similar superficial measurement of three samples at the base of an ice-fall on the uppermost part of Storglaciären showed the crystal size to be approximately 1.8-2.3 cm².

A sample was taken of the layer covering the glacier and formed by refreezing melt water which had percolated through the last winter snow cover. The point was situated 600 m from the glacier margin, 2.5 km from its source and 200 m from Stake No 28. In the first 5 cm of the layer the crystal size was less than 1 mm² and had more the character of frozen firn than real ice. The next 5 cm down contained a mixture of crystals, some of whose areas were less than 1 mm², the larger ones ranging up to 4 mm². This section also had the appearance of frozen firn rather than real ice. From 10 to 20 cm depth the mean size of the crystals was from 4 to 24 mm². In this zone the frozen mass began to take the characteristics of real ice. The next 10 cm of the new ice contained a mixture of crystals varying from very small sizes up to areas of 50 to 100 mm². This zone must be characterized as real ice. Below the dividing line the old glacier ice from 29 cm. down to 32 cm. had a mixture of crystals of about 25 mm² to 200 mm² in area. Then fol-

lowed a band 10 mm thick made up of very small crystals. From this band down to the bottom of the profile, a distance of about 37 cm, three samples of ice were stained and 63 crystal measurements were made from recorded impression. The average size was 1.8 cm^2 and the largest crystal measured 14.0 cm^2 .

An excellent impression was obtained from the cliff on the south side of Kebnepakteglaciären (1.75 km long) in Tarfalsjön (Tarfala Lake) (Fig. 6). Above the cliff the glacier is very steep and nearly forms an ice fall. The regular hexagonal crystals were found to be long and thin, about $4 \times 1 \times 1 \text{ cm}$, most of them with the longest axis perpendicular to the cliff wall. In a long vertical band they were parallel to the wall and in continuation of this band there were three large crystals probably forming the filling to a fissure. The total impression measured was 450 cm and contained about 2050 crystals; Fig. 6 reproduces about 1450; their average size varied between 0.2 and 0.9 cm^2 , the three largest having an average size of 11.0 cm^2 . This unique crystal pattern and structure represent



Fig. 7. Crystal pattern of dead ice in Kebnepakteglaciären

a radical departure from the other type. It is not certain, however, whether it is the real structure of the glacier ice in the cliff or a relatively young superficial layer formed by recrystallization of snow, accumulated on the lake ice and covering the cliff during most of the year.

The latter supposition is strengthened by the fact that the walls of a crevasse in the ice-fall of Isfallsglaciären were covered by a thin layer of regular small crystals measuring about $1\text{--}4 \text{ mm}^2$. Inside it the proper glacier ice consisted of a mixture of crystals, the average size of which was 1.8 cm^2 .

A superficial measurement was made on the dead ice in the lateral moraine on the southern side of Storglaciären where the ice is covered by about 1 m of thick till. The average crystal area was found to be about 11 cm^2 .

Another specimen of dead ice was taken from the northern lateral moraine of Kebnepakteglaciären. The moraine ridge rises up to as much as 30 m in height but the moraine mantle is only a few metres thick. The dead ice must be at least fifteen years old. The sample (Fig. 7) had a very irregular structure. Zone (a) contains only 5 crystals, none of which has a complete boundary intact. One crystal surface is the largest yet measured at Kebnekajse, being 91 cm^2 in area. Three of the crystals are striated with thin, regular air bubble bands, parallel in the same crystal but orientated differently in the various crystals. This striation may indicate individual crystallographic orientation. It is believed that zone (b) with its mixture of long crystals and many small ones is a transition area between zones (a) and (c). In this zone (c) the entire assemblage of crystals is very small and averages in area about 1 mm^2 or less. It may contain about 2500 of these small crystals. Zone (d) is also quite different from all the others and is made up of more normal crystals of various sizes.

It is not possible to arrive at a mean value for the crystal size in the dead ice because of the few measurements made and the wide distribution of sizes observed.

It can be noted, however, that the largest ice crystals recorded in Kebnekajse were found in dead glacial ice.

General Results

(1) All investigations show that a great variation of crystal size and distribution exists in all parts of the glacier ice; therefore an average value of the crystal size in any measured area is not representative. Crystal size frequency analysis must be made.

(2) The nearer the glacial margin, the larger the crystal size; this is particularly striking at the stagnant margin of the glacier.

(3) The largest individual crystals are found in dead ice.

(4) These two paragraphs confirm Seligman's statement that the higher the rate of movement of the glacier ice the smaller is the crystal size.

(5) Cross-section impressions of glacier ice layers are arranged in parallel zones in which the crystals are of large or small mean size.

(6) In dead ice bands of very small crystals ($< 1 \text{ mm}^2$) sharply bounded by crystals of larger size are found.

(7) In the upper of the top layer of the glacier small crystals, formed by refreezing melt water which has precolated through the snow cover of the preceding winter, increase rapidly in size downward to the old glacier surface. In 1948 the contact between this new layer and the old ice was sharp and clearly defined in the ablation area, even in the latter part of the summer.



COMMISSION DE GLACES ET NEIGES

V. — GLACIER MEASUREMENTS

PERMANENT COMMITTEE ON GLACIER MEASUREMENTS

General Report

by

J. M. WORDIE

The Report this year is the first to be made by the new Permanent Committee set up at Washington in 1939 to replace the former Commission Internationale des Glaciers founded by Professor Forel at the Geological Congress in 1894.

The various reports which are now presented fall into two groups : (a) those for Scandinavia and Iceland, and (b) those of Prof. Mercanton for the Alps — French, Swiss, Italian and Austrian.

(a) That on Iceland prepared by Jón Eyfósson covers the period 1930–31 to 1946–47.

Professor Ahlmann reports on Swedish glaciers for the period 1935–1946.

The reports on Norwegian glaciers have been prepared in part by Professor Werenskiöld, but mainly by Fr Faegri. The latter deal with the period 1935–1940 and also for most of these glaciers for each consecutive year from 1940 to 1947. It will be noticed that measurements have been made on most of the glaciers even during the war period.

(b) Professor Mercanton has again supplied a complete report on Alpine variations for 1935–46. As Reporter I should like to take this opportunity of expressing the appreciation which all glaciologists feel for the persistence and energy which Professor Mercanton has given to this research throughout the years.

Following the precedent of reports of the former Commission, all the observers have made their individual reports in tabular form. The advantage of this is that they are clear, practicable and concise.



1) RAPPORT SUR LES VARIATIONS DES GLACIERS DE 1935 à 1946 [47] (Alpes françaises, suisses, italiennes et autrichiennes)

par

P.-L. MERCANTON

Avant-propos

Le soussigné avait préparé pour la réunion de l'UGGI, à Washington, en 1939, un rapport — la quatrième de la série depuis 1914 — sur les variations des glaciers alpins, scandinaves et islandais. Ce rapport a été reçu par l'Association Internationale d'Hydrologie, et donné à l'impression durant l'hiver 1939/40, à Paris. Malheureusement, le tirage n'ayant pu avoir lieu avant l'occupation de la France, les Allemands se sont emparés de la composition et l'ont détruite. Par fortune, le rapporteur en avait gardé une première épreuve qui lui a permis de rétablir le document pour les années 1935/38 et d'en mettre les données en tête du présent rapport, lequel s'étend ainsi à toute la période 1936-1947. Il ne concerne plus que les glaciers de l'Europe continentale, un autre rapporteur ayant pris la charge des appareils nordiques. Quelques résultats de contrôle obtenus dans les Apennins et les Pyrénées françaises figurent ici pour la première fois.

Comme précédemment, on a groupé les données dans des tableaux où, dans la règle, paraissent les glaciers mesurés annuellement ou à intervalles pluriannuels restreints. Dans certains cas, toutefois, quand la variation n'a pu être mesurée mais quand son sens ne fait aucun doute, le régime est indiqué par son signe, le chiffre manquant étant remplacé par un x. Comme auparavant le signe — marque la décrue, l'absence de signe la crue, un 0 l'état stationnaire. Quand le chiffre se rapporte à un intervalle de plus d'une année entre mensurations le nombre des années est inscrit à droite du chiffre de variation, entre parenthèses, ainsi : (2a.), (3a.), etc.

Il s'agit avant tout, dans ces rapports, de variations de longueur, seules données recueillies en nombre suffisant pour obtenir une vue d'ensemble des changements de régime des glaciers d'un territoire. L'ordonnance et le groupement des chiffres concernant une même chaîne de montagne, les Alpes en particulier, sur la base de la nationalité, semblera à bon droit peu scientifique. Elle se justifie néanmoins quelque peu par les conditions dans lesquelles les données sont obtenues. En France elles sont recueillies par l'Administration et revêtent le caractère de documents officiels ; en Italie, c'est la « Comitato Glaciologico Italiano » qui en assume la charge ; en Suisse, c'est la Commission des Glaciers de la Société helvétique des Sciences Naturelles, par les soins des personnels forestiers cantonaux essentiellement. Les glaciers autrichiens sont contrôlés par un ensemble de glaciologues appartenant aux milieux alpinistes allemands et autrichiens. Le détail des mensurations a été publié année après année, en Suisse par la Revue du Club Alpin « Les Alpes » ; en Italie dans le « Bolletino del Comitato Glaciologico Italiano » ; en Autriche par la revue « Zeitschrift für Gletscherkunde ». Les divers collaborateurs tenant à ce que leur travail apparaisse dans son cadre national ce désir a donc été respecté ici. Quand des confusions pouvaient se produire dans la nomenclature (Alpes Orientales), on a donné les noms des glaciers dans les deux langues, italienne et allemande. Quand, dans un massif, les noms usuels de glaciers, ghiacciaio, Gletscher, sont moins usités qu'un terme régional, Vedretta, Ferner, Kees, etc. on a mis le terme local en sous-titre.

Les variations de longueur des appareils sont uniformément données en *mètres*, au demi-mètre près, approximation amplement suffisante et d'ailleurs souvent illusoire déjà.

Il n'est malheureusement guère possible de citer à tout coup les noms de tous les collaborateurs à l'œuvre commune ; on ne nous les a pas toujours communiqués. On nous pardonnera donc d'inévitables omissions. L'avant-propos du Rapport sur la période 1935-38 disait ceci, qu'il sied de rappeler :

« Pour les Alpes françaises c'est la Direction Générale des Eaux et Forêts et en Savoie plus particulièrement le Service du Reboisement de la 5^e Conservation, qu'il convient de mentionner. En Italie, tout le travail, fort diligent, incombe au Comi-

tato Glaciologico Italiano, dont le très actif secrétaire est Mr. Monterin. Ses collaborateurs sont nombreux. Citons, — en nous excusant d'avance en cas d'ommission involontaire — : MM. Videbotti, Camoletto, Vanni, Pignatelli, Nangeroni, Tasa, Capello, Ricci, Conci, Celli, Alby, Brezzi, Peretti, Desio, Ducci, Salocchi, Magnani, Borghi, Mazotta, Faidutti.

Pour les glaciers suisses citons : MM. Jost, Lütshg, Oechslin et Renaud, de la Commission ; puis MM. Campiche, Guex, Maag, Meisser, Streiff-Becker, Vogt, Imhof, Volken, Ambühl, enfin Flotron, pour la Cie des Forces Motrices de l'Oberhasli, avec le professeur Mercanton comme rapporteur régulier.

Les observations des glaciers autrichiens, rassemblés par Mr. le professeur von Klebelsberg, d'Innsbruck, paraissent dans le Rapport sur les mensurations et observations glaciologiques du D u Oe Alpenverein pour la « Zeitschrift für Gletscherkunde ». On y lira les noms de MM. Krasser, Sander, Thalhammer, Leutel, Schatz, von Srbik, Reiner, Paschinger, Heissel, Fresacher ».

Depuis l'établissement du Rapport général pour 1935/38 présenté à Washington, de nombreux changements sont intervenus dans l'organisation comme dans l'exécution des contrôles glaciaires nationaux. Les hostilités ont obligé partout à les restreindre et bien souvent à les abandonner totalement ; il faut regarder comme quasi miraculeux qu'on ait pu, grâce à la ténacité de certains, recueillir encore tant de données durant les années de guerre.

En Italie on a eu à déplorer la mort prématurée du professeur Monterin, l'un des animateurs les plus zélés de ces recherches. Les données italiennes, de 1939 à 1947, ont été rassemblées par Mr. Vanni, privat-docent à l'Université de Turin. Elles émanent de ses propres efforts sur le terrain, comme de ceux de MM. Camoletto, Peretti, Monterin, Pracchi, Padula, Nangeroni, Pignatelli, Riva, Desio, Morandini, Villa, Ricci et Tonini.

Les contrôles autrichiens eux aussi ont été poursuivis avec beaucoup d'énergie et les lacunes y sont peu nombreuses. Mr. von Klebelsberg, qui a bien voulu cette fois encore rassembler ces valeurs, mentionne, outre la sienne propre, la collaboration de MM. Felkel, Fresacher, Gall, Hubatschek, Kinzl, Krasser, Mayer, Mathis, V. et H.P. Paschinger, Petschacher, Sander, Schatz, Senn, von Srbik, Vietoris et Wannemacher.

En Suisse la mobilisation et le surcroît de besogne imposé aux agents forestiers par le ravitaillement du pays en produits sylvicoles ont réduit aux deux tiers, parfois à moins encore, le nombre des glaciers mesurés. Toutefois, les membres de la Commission Helvétique des Glaciers, MM. Jost, Lütshg, Mercanton, Oechslin, et Renaud notamment, avec le concours de MM. Campiche, Guex, Streiff-Becker, Ambühl et Flotron, se sont efforcés de maintenir le contrôle annuel.

Les observations françaises ont souffert beaucoup de la conjoncture. On trouvera néanmoins ici d'utiles données, que Mr. Oudin, directeur de l'Ecole Nationale des Forêts, à Nancy, a pris la peine de rassembler et d'ordonner. Son tableau ne comporte pas uniquement des chiffres de variations de longueur, mais parfois aussi des indications sur les changements de niveau de certains profils transversaux du glacier. Ces variations-là n'impliquent pas nécessairement un sens identique pour la variation de longueur de l'appareil, mais on peut présumer cette concordance et en déduire le régime du glacier. C'est pourquoi on les a données ici.

Le rédacteur du présent rapport général, comme d'ailleurs pour les précédents, ne saurait assumer la responsabilité des données autres que celles des glaciers suisses dont il a le soin spécial. Les correspondants nationaux gardent la leur. Il s'est borné à coordonner les documents reçus sous une forme aussi claire et pratique que possible et de publication pas trop coûteuse. Les précédents établis par les trois premiers rapports de l'ancienne « Commission glaciologique internationale » justifient qu'on conserve leur manière. Le nombre des glaciers européens soumis à un contrôle suivi, annuel le plus souvent, permet en effet une telle présentation statistique. Mais on pourrait au besoin en atténuer la sécheresse inévitable par des compléments descriptifs, illustrés même. C'est d'ailleurs le mode qui s'impose pour mainte région du globe où les contrôles glaciaires sont sporadiques et d'intervalles irréguliers. Mais quel que soit le parti pris définitivement, il importera alors qu'on s'y tienne et que la surveillance des glaciers dans le monde entier soit une œuvre ordonnée et coordonnée nettement.

Pour la Commission des Neiges et Glaciers de l'Association Internationale d'Hydrologie Scientifique :

P.-L. MERCANTON
Lausanne — Suisse
Mai 1948.

Variations de longueur des glaciers des Alpes et Pyrénées françaises

Pour les fronts le signe - indique la décrue, l'absence de signe la crue ; 0 = stationnaire
 Pour les profils absence de signe = exhaussement, — abaissement.

ALPES (Haute Savoie, Massif du Mont-Blanc)

	1935-36	1936-37	1937-38	1938-39	1939-40
Tour (front)	-35 (2a)	—	—	—	—
Argentières (front)	-13	—	-14 (2a)	—	—
Bois (Mer de Glace)(front)	—	-64 (2a)	-17	—	—
Bois (Mer de Glace) profil supérieur	—	- 0,5	-1	—	—
Bossons (front)	35	-42	-35	—	—
» (profil sup.)	—	- 6,5	11,5	—	—
» (profil. inf.)	—	13	5,5	—	—
Bionnasay (front)	-17 (2a)	—	-5 (2a)	—	—
» (profil sup.)	—	—	1	—	—
» (profil. inf.)	—	—	0	—	—
Taconnaz (front)	-4	X	—	—	—
» (profil sup.)	—	-2 (2a)	—	—	—
Tré-la-Tête (front)	—	-X	-7 (3a)	—	—
» (profil sup.)	—	2	-1	—	—
» (profil. inf)	—	- 3 (2a)	-1	—	—

	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Alpes (Savoie).												
Tarentaise												
Sources de l'Isère (front)												
La Vache, (alias La Galise)												
Gébroulaz (front)		-25(2a)		-7,(2a)			-24,5(5a)	-22		-9,5(2a)	2	
» (profil sup.) ..		-15		-1,5(2a)			-3,5(3a)	-15		-5,5(2a)	0	
» (» inf.).....		-1,5		-1(2a)			-16(3a)	0		-4(2a)	3,5	
Maurienne :		0									50 ?	
Sources de l'Are (front)		-11	-24				-x(4a)	-x			-2,5	
» (profil sup.)		-0,5	-1,5				1(4a)	-1,5		-12(2a)	2,5	
» (prof. inf.)		-0,5(2a)	-0,5				6(4a)	0		-3(2a)		
Mulinet (front)	-6(2a)		-40(2a)					-23(5a)		-5(2a)		
Grand Mean (front)			-10(3a)					-65(5a)		-75(2a)		
Evettes (front)	-10		-18					-14		-13(2a)	-x	
» (profil sup.)		-10	-2				-11(4a)	-2		-3,5(2a)	-4,5	
Arnès (front)		-25(2a)					-7,5(4a)			-180(8a)		
Alpes (Isère) Pelvoux												
Pilat (front)			-17,5(5a)							-93(7a)		
» (profil sup.)		2,5	-2,5							-9(7a)		
» (» inf.)		-2	0							-11,5(7a)		
Chardon (front)		0(2a)	-6,5							-100 a-120	(12 a)	
» (profil sup.)		-1,5(2a)	-6							1,5(7a)		
» (» inf.)										-2,5(7a)		
Grandes Rousses :												
La Selle (front)			-20(4a)									
» (profil sup.)			-0,5(4a)									
» (» inf.)			-6,5(4a)									
St. Sorlin d'Arves (front)			-80(4a)							-22(2a)		

France 3

1935-36 1936-37 1937-38 1938 à 1945

Quirlies (front)
Les Malattres (front)

— -30 (4a) — —
— 15 (6a) — —

Pelvoux :

Blanc (front)
» (profil sup.)
» (profil inf.)
Noir (front)
» (profil sup.)
» (profil inf.)

— — - 8 (3a) -28,5 (7a)
— — 2,5 (3a) - 5,7 (7a)
— — 0 - 8,5 (7a)
41 (5a) — 10 (3a) 0
— — 2 (3a) - 2 (7a)
— — - 2,5 (3a) - 6,5 (7a)

Glaciers des Pyrénées :

Pic-Long.

Lac Tourrat..

1906 à 1934- 3 hectares; 1934 à 1945 sensiblement stationnaire

Pays Baché

1896 à 1900- 210 mètres; 1906 à 1937-90 m.

Suite du Bassin du RHONE :	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Ferpècle	- 2,5	- 4,5	—	- 15,5(2a)	- 7,5	- 5,5	- 13,5	- 9	—	- 18(2a)	31,5	- 21,5
Arolla	- 4	- 10,5	—	- 26(2a)	- 19	- 3	22,5	- 21	—	- 37(2a)	- 16	- 23
Tsigliorenoue	- 9	—	—	- 42(3a)	- 24	0	- 25,5	15	—	- 26(2a)	20	0
Duran (Seillon)	- 5	- 19	- 13,5	—	- 7(2a)	- 3	- 8,5	- 2,5	11,5	- 2	- 2	- 9
Lendarrey	- 6,5	- 3	- 10	—	- 3(2a)	- 6	- 4	- 9,5	- 13	- 7,5	- 5,5	- 5
Grand Désert	- 10,5	—	- 15	—	- 5,5(2a)	- 14	- 32,5	23,5	- 21	- 17,5	- 4,5	- 27,5
Mont Fort	- 1,5	—	- 5	—	- 3,5(2a)	- 1	- 9,5	12,5	- 11	- 11,5	- 1	- 13
Valsorey	- 7,5	0	0	- 60	- X	- 4	- 14,5	- 14	- 25	- 43,5	—	- 9,5
Tseudet	—	—	—	—	—	- X	- X	- X	—	—	—	—
Salemaz	- 10	- 14	- 11	- 23,5	- 23,5	- 5	- 26	- 6	- 2,5	0	- 12	- 32
Trient	- 20	- 12	- 25	- 14,5	- 21	- 6	- 6	- 3	- 3	- 15	- 16	- 17
Grand Plan Nèvé	—	20(2a)	- 13	- 16,5	—	- 11(2a)	—	41(2a)	—	—	- 60(3a)	- 36
Petit Plan Nèvé	—	0(3a)	0	- 1,5	—	- 34(2a)	—	21(2a)	—	—	- 30(3a)	- 18
Martinets	—	8,5(2a)	- 2	- 2	—	- 6(2a)	—	0	—	—	- 46(3a)	- 13
Paneyrossaz	—	- 31(2a)	- 26,5	—	—	45(3a)	—	—	—	—	—	331(6a)
Prapioz	0	0	- 13	- 4	- 5	- 6	- 47	- X	- 58(2a)	—	64(2a)	- 88
Scex Rouge	—	5(2a)	10	0	- X	—	—	—	- 90(3a)	—	87,5(2a)	- 91

Bassin de l'Aar : 1935-36 1936-37 1937-38 1938-39 1939-40 1940-41 1941-42 1942-43 1943-44 1944-45 1945-46 1946-47

Oberaar	-18,5	-20	-16,5	-14	-11,5	-7	-21,5	-18	-22	-13	-20	-22,5
Unteraar	-41	-40	-30	-26	-13,5	-26	-11,5	-29	-20	-10	-10,5	-17
Trift	—	—	-X	-X	-X	—	—	—	—	—	—	—
Rosenlau	-10	-11	-8	-X	-X	-X	-38(4a)	-3,5	-X	-X	-6	-X(7a)
Grindelwald Supérieur	-14	-17	-30,5	-X	-X	-X	-10	-10	-6	-20	-25	-23
Grindelwald Inférieur	-2	-X	-X	-2	-X	-X	—	—	—	—	—	-32
Eiger	—	-18(2a)	-8	—	—	—	—	—	—	—	—	—
Stein	0	-10	-29	-43	-11	-4	-3	0	0	0	0	0
Blümhisalp	-7	-10	-7	—	-17(2a)	0	-7	-12	—	-5,5(2a)	-14	-7
Schwarz	3	1,5	-2	10	—	16,5(2a)	4	—	0(2a)	—	-3(2a)	-1,5
Rätzli	-7	-10	-25,5	-3,5	—	-3,5(2a)	-16	-26	—	—	-28(3a)	-34
Tschingel	—	—	-28,5(3a)	—	—	—	—	—	—	—	—	—
Ganchi	—	—	-10,5(3a)	—	—	20(3a)	—	—	-1(3a)	-1	-3,5	—
Balmhorn	—	—	-15(3a)	—	—	—	—	—	—	—	—	—
Tsanfleuron	1	-26	-10	-6	-5	1	-11	-14	-33	-3	-10	-41
Wildhorn	—	—	—	—	—	-11(2a)	—	-16(2a)	—	—	-10(3a)	—

[illegible]

Bassin de la LINTH :

Sulz	- 5,5	- 6	- 2,5	- 9(2a)	6,5	- 9,5(2a)	X	0	- 6,5
Gärnisch	- 18(2a)	—	- 5(2a)	—	—	- 14(2a)	—	- 3(2a)	—
Clarides	- X	—	- 5(2a)	—	X	- X	—	—	—
fünften	- 4	—	- 12(2a)	—	—	—	- X	X	—

<i>Bassin du RHIN :</i>	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Piz Sol	—	8	—	—	—	-36(3a)	-12,5	—	—	2,5(2a)	1,5	X
Sardona	—	-X	—	—	—	3(4a)	—	—	—	-133(4a)	-10	-25
Punteglas	3	-3	—	—	-18,5(2a)-16	-9	—	—	-14(2a)	7	-2	-42,5(2a)
Vorab	2,5	-4,5	3,5	-4	—	-1,5(2a)-8	-8	-47,5	8,5	—	-2	-33
Lorch	7	-7	—	—	-22(2a)	-18	-14,5	-23,5	5	—	—	-22,5
Porchabella	—	-11,5	-11	—	-16(2a)	7,5	—	-23(2a)	—	10(2a)	-28	-41
Paradies	—	-28(2a)	-45	—	-17,5(3a)-20	-17,5(3a)-20	-39,5	-39,5	—	23(2a)	-14	-46
Verstankla	-22(2a)	-7,5	-10,5	—	-13(2a)	-2	-10	-8,5	-25,5	9	-3,5	-56
Lenta	-1,5	6	—	—	-4(3a)	-2	-10	-2,5(2a)	—	8(2a)	0	-19
Schwarzhorn	—	-5,5	-5,5	—	—	-10(3a)	—	-14,5	—	-17(2a)	-2	-17,5
Suretta	—	—	—	—	—	—	—	—	—	—	—	—
<i>Bassin de l'INN:</i>												
Mortersatsch	-7	-12,5	-25,5	-34,5	-42	-22,5	-22,5	-20	-34,5	-21,5	-19,5	-47,5
Roseng	-15	-9,5	-8,5	-23,5	-11,5	-17,5	-8,5	-12,5	X	-26	-14	-12,5
Tschierwa	—	—	—	—	—	—	—	—	—	9	-12	-13
Tiatscha	—	—	-7,5(3a)	—	—	13(2a)	-32,5	-24	2	-51	-9	-30,5
Lichanna	—	—	—	—	-18(5a)	-18(5a)	-16(2a)	—	—	6	0	-17,5
Piquogl	—	-22(2a)	—	—	—	-29(4a)	-10	-14	-3,5	-11	-12	-13
<i>Bassin de l'ADDA :</i>												
Forno	—	-17(2a)	—	34,5(2a)-42	-42	—	—	-36,5(2a)-18,5	—	-23(2a)	-19	-30,5
Cantone	—	—	-21(3a)	—	-17(2a)	—	—	—	—	—	—	—
Albigna	—	-X	-X(3a)	—	-X(2a)	—	—	—	—	—	—	—
<i>Bassin de TESSIN :</i>												
Rossboden	-16,5	-51,5	-119	—	—	15	12,5	-16	-10	—	8,5	-1,5
Basodino	—	17(2a)	0	15	19	—	-53(2a)	—	-12(2a)	1	-19	-116
Bresciana	—	-X	-6,5	2,5	0	11	-3	-7	-2,5	-5,5	—	-105(2a)
<i>Récapitulation % de Glaciers en décroissance</i>	9	12	4	14	15	24	6	7	7	8	10	0
	84	76	83	81	78	72	94	89	87	87	80	98

Variations de longueur des glaciers des Alpes italiennes (en mètres)
Glaciers (ghiacciaio, Vedretta)

(le signe- indique la décrue; l'absence de signe, la crue; 0, signifie stationnaire; a, année)

Bassin du PO :

I. — Groupe des Alpes Maritimes (Gelas, Maledia, Argentera)

	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46
Clapier	6,5	- 2	- 4	3	—	X(2a)	- 2	—	—	—	—
Petrabroc	2	- 2	-15	0,5	—	4(2a)	- 3	—	—	—	—
Maledia	2,5	- 0,5	1	0,5	—	5(2a)	- 4	—	—	—	—
Muraion	7,5	- 2	8	3,5	—	6(2a)	- 8	—	—	—	—
Gelas E.	5,5	- 2,5	- 0,5	3	—	X(2a)	- 2	—	—	—	—
Gelas NE	1,5	- 1	- 0,5	2,5	—	X(2a)	- 2	—	—	—	—
Gelas N.	-3	- 2	1	3,5	—	X(2a)	- 2,5	—	—	—	—
De Cessole	2,5	- 2	3	4,5	—	—	—	—	—	—	—

II. — Groupe Cenisio-Ambin (Dora Riparia)

Fourneaux	4	-10	- 8	—	—	—	—	—	—	—	—
Galambra	0	- 9,5	- 9	—	—	- 6	—	—	—	—	—
Agnello S.	0	-11	- 4	—	—	-12	—	—	—	—	—
Agnello W.	-X	-X	-X	—	—	-X	—	—	—	—	—
Agnello E.	—	—	- 1	—	—	-350(3a)	—	—	—	—	—
Muttett	—	—	-X	—	—	4	—	—	—	—	—

III. — Groupe Gran-Paradiso (Valle dell'Orco e Dora Baltea)

Valetta	—	—	—	-12	—	-X	—	—	—	—	—
Patri inf.	4	- 6	- 5	- 3	—	3	—	—	—	—	—
Coupé di Money	1	- 3	-10	- 2	—	1	—	—	—	—	—
Money	- 2,5	- 8	- 4,5	- 4	—	4	—	—	—	—	—
Grand Croux	0	-12	-11	-X	—	13,5	—	—	—	—	—
Tribolazione	40	-16	-X	20	—	-20	—	—	—	—	—
Valeille	-20	-15	-15	- 4	—	6	—	—	—	—	—
Herbetet	—	—	—	- 6	—	—	—	—	—	—	—
Sengie	0	0	- 8,5	0	0	X	—	—	—	—	—

IV. — Groupe Mont Blanc

1935-36 1936-37 1937-38 1938-39 1939-40 1940-41 1941-42 1942-43 1943-44 1944-45 1945-46

Pré de Bar	-20	25	-22	-25	-20	X	-X	-8	—	—	-X
Triplet	-25	-40	-35	-50	-10	X	-X	0	—	—	-X
Miage	0	-10	-6	—	—	0	0	—	—	—	—
Allée Blanche	-25	0	-10	-3	-5	0	-X	-5	—	—	—

ALPES PENNINES :

1. Valpelline.

Tsa de Tsan	—	-9	-19	-5	-31	—	—	—	—	—	—
Morion	—	—	0	—	-6	—	—	—	—	—	—
Florio	—	—	-X	—	-X	—	—	—	—	—	—
Des Dames	—	—	—	—	-7(2a)	—	—	—	—	—	—

2. Valtournanche

Mont Tabel	—	—	-X	-X	—	—	X	—	—	—	—
Chérillon	-1,5	-1,5	-7	1,5	-3	3	-7	-9	-5	-3	2,5
Tyndall	—	—	-8	-6,5	X	7,5	-6	-10	—	-19(2a)	-9,5
Cervino	—	-5	0	-X	10	-3	-15,5	-3,5	—	-12(2a)	-9,5
Valtournanche	—	-14	-5	-X	-X	-24	-12,5	-19,5	-10	-20	-10
Teodulo Inf.	—	5	0	-X	—	-X	—	—	—	—	—

3. Monte Rosa

Grande Verra	-13,5	-19	-11,5	-24	-7,5	-5	-8,5	—	—	—	—
Piccolo Verra	0	-8	-10,5	-20	-0,5	—	-23(2a)	—	—	—	—
Lys	8,5	-20	-20,5	-14,5	-17	-48,5	-4,5	-33	—	-1,5(2a)	-16
Indren W	—	—	-14(3a)	0	-1,5	-5	-6	—	—	—	-X
Sesia	—	0	-5	1	29,5	X	-X	—	—	—	-7,5
Bors	—	-1,5(2a)	0	—	—	—	—	—	—	—	—
Belvedere	—	-17,5	-22	-5,5	3	-7,5	-3,5	—	—	—	—
Filar	—	—	13,5	13	3,5	—	0	—	—	—	—
Perazzi	-5	—	-41(2a)	—	—	—	—	—	—	—	—
Piode	-8	—	-45,5(2a)	—	—	—	—	—	—	—	—
Castore	0	—	-23(2a)	—	—	—	—	—	—	—	—

ALPI RETICHE
Bassin de l'ADDA :

1. — Groupe Tambo, Suretta, Stella-Ligoncio.		1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46
Valle Loga	—	—	—	-21(3a)	—	-9(2a)	—	—	—	—	—	-17(6a)
Pizzo Ferrè	- 2	- 1	- 4	- 4	—	-6(2a)	—	- 3(2a)	—	—	—	-21(4a)
Spianata	0	- 3	- 5	- 5	—	-6,5(2a)	0	—	—	—	—	—
Suretta	—	-11	-13	-13	—	-20(2a)	- 4	—	—	—	—	—
Orsareigis	-11(2a)	- 4	- 6	- 6	—	- 8(2a)	—	—	—	—	—	—
Calvo	—	- 5	- 6,5	- 6,5	—	—	-29(3a)	—	—	—	—	—
Badiletto	—	- 8	- 7	- 7	—	-11(2a)	- 6	—	—	—	—	—
Badile	—	- 7	- 9	- 9	—	-13(2a)	- 7	—	—	—	—	—
Cengalo SW	—	- 6	- 8	- 8	—	-15(2a)	- 5	—	—	—	—	—
Cengalo SE	—	- 9	-12	-12	—	-19(2a)	- 9	—	—	—	—	—
Ferro W.	—	-84(17a)	-10,5	-10,5	—	-4(2a)	—	—	—	—	—	—
Ferro Centrale	—	-88(17a)	-11	-11	—	- 5(2a)	—	—	—	—	—	—
Poncigna	0	- 3	- 1	- 1	—	—	—	—	—	—	—	-93(3a)
Scalino	—	—	—	—	—	—	—	—	—	—	—	—

2. — Groupe Disgrazia-Bernina.

Cassandra	- 7	- 8,5	- 5	—	—	—	—	—	—	—	—	—
Ventina	- 6	-16	-15	- 3	- 8	—	—	—	—	—	—	—
Disgrazia	-33	-24	-31	45	49	—	—	—	—	—	—	—
Cima di Rosso	- 5	—	- 2	—	—	—	—	—	—	—	—	—
Scerscen inf.	—	—	-X	—	—	—	—	—	—	—	—	-35(7a)
Scerscen sup.	—	—	-X	-15,5(7a)	—	—	—	—	—	—	—	—
Caspoggio	- 8	-15	-18	11	—	—	- 7(2a)	—	—	—	—	—

III. — *Groupe Alpes Orobie.*

	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-49
Lago Malgina	—	7(3a)	7	—	—	—	—	—	—	—	—
Trobio	—	-10(3a)	—	—	—	—	—	—	—	—	—
Porola	0	1	—	—	—	—	—	—	—	—	—
Pizzo Rodes	—	3(2a)	—	—	—	—	—	—	—	—	—
Lupo	0,5	1	2	—	—	—	—	—	—	—	—
Marovin	—	-10(4a)	-11	—	—	—	—	—	—	—	—
Druito	—	-1(2a)	-26	—	—	—	—	—	—	—	—
Cagamei I	—	—	—	1	-1	—	—	—	—	—	—
Cagamei II	0	0,5	4	2	-2	—	—	—	—	—	—
Cagamei III	0	1	-18	2	-2	—	—	—	—	—	—
Bondone	—	—	-40(4a)	—	—	—	—	—	—	—	—

Bassin de l'ADDA : Groupe Ortler-Cevedale.

Confinale SW	—	—	-1,5(3a)	-1,5	0	—	—	—	—	—	—
Confinale SE	—	—	-14(3a)	-14	—	—	—	—	—	—	—
Cime dei Forni	—	-79(2a)	7	7	—	—	—	—	—	—	—
Rosole	7	7,5	-25	-25	0	—	-24	—	—	—	-15,5(5a)
San Giacomo E	—	0	9	8	0	—	—	—	—	—	—
Dei Forni	—	-79(2a)	-12	-30,5	—	—	-41	—	—	—	-33,5(5a)
Cerena	—	-26(2a)	53	—	—	-118	—	—	—	—	—
Sobretta NE	—	—	-13,5(3a)	—	—	—	—	—	—	—	—
Platighole	-19,5	0	—	-6,5(2a)	—	—	—	—	—	—	—
Dei Vitelli	-18,5	—	—	-24(3a)	—	14	—	—	—	—	—
Zebru	-16,5	-10	—	-18(2a)	—	—	—	—	—	—	—
Della Miniera	-45,5(2a)	—	—	—	—	—	—	—	—	—	—
Dei Castelli E'	7(2a)	—	—	-5(3a)	—	—	—	—	—	—	-8,5
Dei Castelli W	-13(2a)	X	—	-28,5(3a)	—	—	—	—	—	—	-13,5

<i>Bassin de l'ADDA (suite)</i>	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46
Montagna Vecchia	—	—	—	—	—	—	—	—	—	—	- 2,5
Del Fora W	- 5(2a)	—	—	—	—	—	—	—	—	—	—
Del Fora E	- 2(2a)	- 17	—	—	—	—	—	—	—	—	—
Gran Zebro	—	- 37(2a)	- 32	- 8,5	—	- 9,5(2a)	—	—	—	—	- 25 (5a)
Cedé	—	- 27(2a)	- 4,5	- 5,5	—	9 (2a)	—	—	—	—	- 10 (5a)
Isola Persa	—	- 3,5(2a)	—	—	—	—	—	—	—	—	—
Pasquale N	—	—	- 2(3a)	- 2,5	—	—	—	—	—	—	—
Pasquale S	—	—	- 6(8a)	—	—	—	—	—	—	—	—
Dosegu	—	—	- 127(3a)	—	—	- 50,5(3a)	—	—	—	—	—
Passo del Dosegu	—	—	- 20(3a)	—	—	- 14 (3a)	—	—	—	—	—
Punta Storzellina N	—	—	- 11(3a)	—	—	- 6 (3a)	—	—	- 10 (3a)	—	—
Punta Storzellina W	—	—	0(3a)	—	—	- 7 (3a)	—	—	1 (3a)	—	—
Sforzellina	—	—	- 32(3a)	—	—	- 20 (3a)	—	—	- 6,5(3a)	—	—
Lago Bianco	—	—	- 21(3a)	—	—	—	—	—	—	—	—
Gavia	—	—	- 7(4a)	—	—	—	—	—	—	—	—
Val dell' Alpa S	—	—	- 22(4a)	—	—	—	—	—	—	—	—
Val dell' Alpa M	—	—	- 15(4a)	—	—	—	—	—	—	—	—
<i>Bassin de l'ADIGE :</i>											
Zai N	7	—	- 6(2a)	- 6,5	—	—	- 5,5(3a)	—	—	—	—
Zai Centrale	- 9	—	- 11(2a)	- 7	—	- 20,5(2a)	- 7	—	—	—	—
Zai S	1,5	—	- 11(2a)	—	—	- 7 (3a)	- 19	—	—	—	—
Rosim	—	—	- 11(3a)	- 5	—	—	—	—	—	—	—
Madriccio	- 9,5	—	- 7(2a)	- 9	—	- 9,5(2a)	- 20	—	—	—	—
Solda	- 6	—	- 10(2a)	- 1,5	—	- 0,5(2a)	- 9	—	—	—	—
Finimondo	- 4	—	0(2a)	—	—	—	—	—	—	—	—
Marlet N	- 49,5	—	- 18(2a)	—	—	—	—	—	—	—	—
Marlet S	- 16,5	—	- 15(2a)	—	—	—	—	—	—	—	—
Laste	- 21	—	X(2a)	- 11	—	—	—	—	—	—	—
Beltovo	- 14,5	—	—	- 18,5(3a)	—	—	—	—	—	—	—

4. — *Groupe Adamello-Presanella* : Italie 7
1935-36 1936-37 1937-38 1938-39 1939-40 1940-41 1941-42 1942-43 1943-44 1944-45 1945-46

Venerocolo	—	—	—	18,5	—	5	0	0		
Pisgana E	—	—	—	73,5(5a)	—	—	—	28,5(3a)		
Pisgana W	—	—	—	96,5(5a)	17,5	22,5	7	—		
Nardis W	—	—	-20	—	-37(2a)	—	18	—		
Corsinello	—	—	-1,5	—	-42(2a)	—	32	—		
Lobbia	—	—	—	-135(5a)	15	0	0	—		
Forgorida	—	—	—	—	26,5	—	—	—		
Lares	—	—	—	—	10	—	—	—		
Nisch	—	—	—	-83(5a)	—	—	—	—		
Carè Alto E	—	—	—	-112(5a)	-22	—	—	3,5		
Amola	—	—	—	—	11(6a)	—	—	—		
Monte Nero	—	—	—	—	—	210(7a)?	-50	—		
Presanella	—	—	—	—	—	-60	-X	—		
						-100	-38	—		

5. — *Alpes Venoste* :

Vallelunga (Langtaufere)	0	-19	-19	-31	—	24(2a)	—	62	—	-19	—	-35,5(3a)
Barbadorso Centrale	-14	-13,5	-31,5	-11	—	-43(2a)	—	17	—	-4	—	-20,5(3a)
Fontana E (Freibrunner)	-5,5	-19,5	-9,5	—	—	-19(3a)	—	5,5	—	-10	—	-23(3a)
» W	-5,5	-12	-87,5	—	—	-115(3a)	—	15	—	-48	—	-24,5(3a)
Pianolo	-5,5	-5,5	-21	—	—	—	—	-120(4a)	—	-2	—	—
Di Mazia (Matscher)	-5,5	-5,5	-14	-24	—	-111,5(3a)	-46	—	—	—	—	—
Oberettes W	0	-8,5	-3,5	—	—	—	—	9,5(4a)	—	—	—	—
» E	7	-5,5	-1	—	—	—	—	2(4a)	—	—	—	—
Di Saldura	-12	-15,5	8,5	—	—	—	—	2,5(4a)	—	—	—	—
Ramudla	—	11,5(2a)-11	—	—	—	—	—	-70(4a)	—	—	—	—

Suite de	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46
5. — <i>Alpes Venoste</i> :											
Delle Frane (Steinschlag)	- 5,5	59	- 46	—	—	—	- 15	—	—	—	—
Di Gлого Alto	- 3	- 6	- 4,5	—	—	—	—	—	—	—	—
Tessa	- 0,5	- 12	- 12	—	—	—	—	—	—	—	—
Riotorbo W I	- 7	- 21	- 12	—	—	—	—	—	—	—	—
Riotorbo W II	- 3,5	- 8,5	- 3	—	—	—	—	—	—	—	—
Riotorbo N	- 2,5	- 11	- 15,5	—	—	—	—	—	—	—	—
Ceppo	—	- 1(2a)	- 1	—	—	—	—	—	—	—	—
Croda del Cavallo destro	—	8	- 1	—	—	—	—	—	—	—	—
Croda del Cavallo sinistro	- 6,5	11	- 4,5	—	—	—	—	—	—	—	—
Cigot	0	- 6,5	- 13,5	—	—	—	—	—	—	—	—
Quaira	—	15	- 2,5	—	—	—	—	—	—	—	—
Fiammante E	—	2,5	—	—	—	—	—	—	—	—	—
Vacche	2	- 1,5	0	—	—	—	—	—	—	—	—
Centrale del Lago	- 6,5	- 8	- 10,5	—	—	—	—	—	—	—	—
Malavalle Sinistro (Uebeltalferner)	—	- 6,5	1,5	—	—	—	—	—	—	—	—
Malavalle Centrale	- 8	- 18,5	- 17,5	—	—	—	—	—	—	—	—
Malavalle Destro	- 9	—	- 22,5(2a)	—	—	—	—	—	—	—	—
Pendente	- 6	- 5,5	- 6,5	—	—	—	—	—	—	—	—
Stua	- 1,5	—	- 5 (2a)	—	—	—	—	—	—	—	—

Variations de longueur des glaciers des Alpes autrichiennes (en mètres)

(le signe — indique la décrue; l'absence de signe : la crue; 0 signifie stationnaire; a, Année) — Glacier (Ferner, Kees)

DACHSTEIN:	1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Gr. Gosau	-14,5	-13	—	—	-15,5(3a)	-10,5	2	—	-25,5(2a)	—	-60 (2a)	-28
Hellstätter	-8	-6,5	—	—	-19 (3a)	-4,5	-5	—	-27,5(2a)	—	-20,5(2a)	-25
Schladminger	—	-5,5	—	—	—	—	—	—	—	—	—	-4,5

SILVRETТАGRUPPE

Litzner NE	-6,5	-5	-1,5	-4	-1	-2,5	-2	-8,5	-9,5	-3,5	0	-13
Litzner SW	-1,5	-4	-0,5	2	2	-1	0	-5	-2	0	-1,5	-8
Klostertal	-24	-23	-21	-8	-2,5	-4	-22,5	-5,5	-20,5	-12,5	-13	-19
Vermunt W	1	-25,5	-13,5	-13	-10	-31,5	-53	-29,5	-41,5	-39	-34	-18
Vermunt E	-6,5	-10	-11,5	-5	-4	-10,5	-11,5	-1,5	—	-14,5	-14	-22
Bieltaler	-10,5	-17,5	-12	—	—	-16,5	-15,5	-16,5	—	-13	-18	-20
Jamtaler	-10,5	-6,5	-21	—	—	-36	-26	-21	-73	-4,5	-19	-9
Larain	-8	-4,5	-8	—	—	-6,5	-7,5	-24,5	—	-17,5	-6	-9

AUTRICHE 2

OETZTALER ALPEN (Ferner)		1935-36	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Gepatsch	-10,5	-15,5	-12	-	-	-26(2a)	-17	-9	-32	-15	-	-13,5(2a)-	26
Weiss-See	-19	-14	-12	-	-	-	-	-17	-24,5	-14	-	-120 (2a)	-
Sexeggarten	-12	-18,5	-11,5	-	-	-	-	-40,5	-31	-31	-25	-45,7	-27
Paschach	-	-14,5	-16	-	-	-115(2a)	-10	-18	-95	-153	-102	-48	2
Mittelberg	-21	-41,5	-	-	-	-63 (3a)	-5,5	-	-	-27	-2,5	-25,5	-25,5
Hochjoch	-	-	-12,5	-12	-12	-18	-25	-17,5	-	-12,5(2a)-	27,5	-	-14(2a)
Hinterreis	-35	-17,5	-	-27 (2a)-	-66	-10,5(2a)	-8	-54	-32,5	-6	-25	-35	-110
Guslar	-15	-19	-7	-	-	-	-11,5	-	-17	-33	-20	-17	-30
Vernagt	-27,5	-19,5	-	-7 (2a)-	-6	-	-2,5	-22,5	-20	-7,5	-20	-20	-60
Mitterkar	-8	-8,5	-4	-12	-	-	-6	-6	-12,5	-11,5	-3,5	-6	-18
Rofenkar	-21	-27	-10	-9,5	-	-	-13,5	-5,5	-6	-6,5	-6	-11,5	-15
Taufkar	-3	-1,5	-1,5	-13,5	-	-	-29,5	-8	-5,5	-6	-5,5	-5	-61
Niederjoch	-12,5	-21	-11,5	-4,5	-13	-	-11,5	-13,5	-6	-21,5	-14,5	-20	-29
Marzell	-2,5	-11,5	-4,5	-29	-25	-25	-11,5	-8,5	-23	-36,5	-5,5	-69	-35
Schalf	-3,5	-8,5	-15,5	-18	-95	-95	-13,5	-29	-11	-22	-106,5	-21,5	-18
Diem	-2,5	-12,5	-13,5	-7	-19	-19	-8,5	-21,5	-24	-28	-17	-22	-13,5
Spiegel	-9	-9	-10	-11	-	-8,5	-9	-17	-16,5	-16	-15	-7,5	-17,5
Gurgler	-	-	-	-	-	-	-	-8,5	-25,5	-19,5	-12	-10	-13
Langtaler	-8	-	-33 (2a)	-	-	-33 (2a)	-19	-17,5	-16	-20,5	-15,5	-16,5	-17,5
Rotmoos	-17	-	-46,5(2a)	-	-	-18,5(2a)	-13,5	-18,5	-27	-21,5	-37,5	-11,5	-25
Gaisberg	-4	-	-16 (2a)	-	-	-11 (2a)	-6	-5	-7	-16,5	-	-16,5	-10,5

STUBAIER ALPEN (Ferner)

	1935	36	1936	37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Sulztal	—	2 (2a)	—	—	23 (2a)	—	-37,5(2a)	—	-15,5(2a)	—	-35 (2a)	—	-44 (2a)	—
Bockkogel	—	-11,5(2a)	—	—	48,5(2a)	—	-16,5(2a)	—	—	—	-137 (4a)	—	-19 (2a)	—
Schwarzenberg	—	-23 (2a)	—	—	-39,5 (2a)	—	-38 (2a)	—	—	—	-130 (4a)	—	-58 (2a)	—
Bachfallen	—	-1,5	—	—	-10 (2a)	—	—	—	10 (4a)	—	-21 (2a)	—	-24,5(2a)	—
Winnebach	—	0 (2a)	—	—	-10 (2a)	—	—	—	-20 (4a)	—	X (2a)	—	-10 (2a)	—
Längentaler	—	-9 (2a)	—	—	-17 (2a)	—	—	—	0 (2a)	—	-10,5(2a)	—	-11,5(2a)	—
Lisenser	—	—	—	—	—	—	-27 (2a)	—	-21,5(2a)	—	-37 (2a)	—	-40 (2a)	—
Alpeiner	—	-60 (2a)	—	—	-52,5(2a)	—	—	—	24 (4a)	—	-19,5(2a)	-9	-22	—
Berglas	—	-6,5(2a)	—	—	-15 (2a)	—	-13,5(2a)	—	-23 (2a)	—	0 (2a)	-0,5	-8	—
Daunkogl	—	8	—	12,5	-28	-13,5	—	—	-11 (2a)	—	-30(2a)	-18	-15	-48,5
Schaufel	—	1,5	—	21,5	-41,5	-0,5	-8,5	—	-7(2a)	—	-10,5(2a)	-16,5	-24	-37,5
Fernau	—	1,5	—	11	-54,5	-36	-1,5	—	X (2a)	—	-17,5(2a)	-28	-0,5	-6
Sulzenau	—	-5	—	33	-43	—	-31 (2a)	—	-23,5(2a)	—	-9,5(2a)	-39,5	-23	-34
Grünau	—	-13,5	—	24	-15	-8	-14	—	-14 (2a)	—	-10,5(2a)	-10,5	-12,5	-11
Grübl W	—	0	—	5,5	-20,5	—	—	—	-5 (4a)	—	-21,5(2a)	-18	-11	-21,5
Grübl E	—	2	—	7	0	—	—	—	—	—	—	—	-12 (2a)	-14,5
Simming	—	-7,5	—	-136	-11,5	-7,5	—	—	-6,5(3a)	—	-7 (2a)	-13,5	-10	-32

ZILLERTALER ALPEN (Kees)

Waxeegg	-27	-22,5	-40	-27,5	-22	-30	-85	—	-59,5(2a)	—	—	-38,5
Horn	-11	-14,5	-10	-12	-18,5	—	-37,5(2a)	-20	-40,5	-18	-35	-27,5
Schwarzenstein	-7	-22	-13	-11	-8,5	—	-28,5(2a)	-19	-19	-18	-27	-28
Schlegeis	—	-X	-X	—	—	—	—	—	—	—	—	—
Furtschagel	-18,5(2a)	—	-31 (2a)	—	—	—	—	—	—	—	—	—

VENEDIGERGRUPPE (Kees)
 1935-36

	1936-37	1937-38	1938-39	1939-40	1940-41	1941-42	1942-43	1943-44	1944-45	1945-46	1946-47
Krimmler Mittl.	—	—	-46 (2a)	—	-113 (2a)	—	-93 (2a)	—	—	-79 (3a)	—
Krimmler E.	-21,5(2a)	—	-26 (2a)	—	—	—	-59 (3a)	—	—	-5 (3a)	—
Obersulzbach	-21,5(2a)	—	-16,5(2a)	—	-37 (2a)	—	-52 (2a)	—	—	-152(3a)	—
Untersulzbach	-33 (2a)	—	-29 (2a)	—	-31 (2a)	—	-54 (2a)	—	—	-49 (3a)	—
Habach	-33 (2a)	—	-7 (2a)	—	—	—	-38,5(4a)	—	—	-59,5(3a)	—
Umbal	-29 (2a)	—	-42,5(2a)	—	1 (2a)	—	-113 (2a)	—	—	-36 (3a)	—
Simony	-12,5(2a)	—	-32 (2a)	—	-70,5(2a)	—	-17,5(2a)	—	—	-22 (3a)	—
Maurer	-15 (2a)	—	-37,5(2a)	—	-20,5(2a)	—	-37 (2a)	—	—	-28,5(3a)	—
Dorfer	-22 (2a)	—	-19 (2a)	—	-26 (2a)	—	-20 (2a)	—	—	-68 (3a)	—
Mullwitz	-32 (2a)	—	-30 (2a)	—	-31 (2a)	—	-28,5(2a)	—	—	-37 (3a)	—
Frossnitz	-25 (2a)	—	-32 (2a)	—	-31,5(2a)	—	-35	—	—	-35 (3a)	—
Schlatten	-14 (2a)	—	-16,5(2a)	—	-28 (2a)	—	-30,5(2a)	—	—	-35,5(3a)	—
Viltragen	-28,5(2a)	—	-9 (2a)	—	-31 (2a)	—	-31,5(2a)	—	—	-39,5(3a)	—

GLOCKNERGRUPPE (Kees) :

Karlinger	-17 (2a)	—	-80,5(2u)	—	-93,1(2a)	—	-52,5(2a)	—	-56 (2a)	—	-68,5(2a)	—
Bärenkopf	-17,5(2a)	—	—	—	—	—	-8,5(2a)	—	-16 (2a)	—	-6,5(2a)	—
Klockerin	-13,5(2a)	—	-12,5(2a)	—	-11 (2a)	—	-5	—	-9	-5	-20,5(2a)	-16
Pasterze	5	-7	-3	-8,5	-2	-5,5	—	-10(2a)	6,5	-125	-125 (2a)	—
Pfandscharten	—	—	—	—	-1,5	—	—	—	13	-52,5	-52,5(2a)	-81
Wasserfall	—	—	—	—	—	—	—	—	—	—	—	—

ANKOGELGRUPPE (Kees) :

Grosselend	-7	-5	-8	-19	-10	-16,5	-20	—	-28 (2a)	—	-100(2a)	-21
Kleinend	-10	-4,5	-9,5	-11,5	-9	-11	-8	—	-19 (2a)	—	-74(2a)	-28
Kälberspitz	—	-8 (2a)	-6	-10	-10	-9,5	-9	—	-24,5(2a)	—	-22,5(2a)	-22
Tripp W.	-4 (2a)	—	-4,5(2a)	-3	—	-3	-2	—	-26(2a)	—	—	—
Hochalm r. Lappen	-7,5	-7,5	-9	-5	-18,5	-12	—	—	—	—	—	—
Winkel	-40,5(5a)	—	-3,5(2a)	-11	-7,5	—	—	—	—	—	-200(7a)	—

RAPPORT SUR LES VARIATIONS DES GLACIERS DE 1930 A 1946/47 (ISLANDE, SUÈDE ET NORVÈGE)

ICELAND

FORANDRINGER VED ISLANDSKE JÖKLER 1930 - 1947 (OG 1904-1940)

	1930	'31	'32	'33	'34	'35	'36	'37	'38	'39	'40	'41	'42	'43	'44	'45	'46	'47
	/31	/32	/33	/34	/35	/36	/37	/38	/39	/40	/41	/42	/43	/44	/45	/46	/47	
I. DRANGAJÖKULL																		
1. Kaldalón																		
2. Leimufjörður																		
3. Reykjafjörður																		
4. Paralátursfjörður																		
II. SNÆFELLSJÖKULL																		
1. Hymningsjökull																		
2. Jökulháls																		
3. Nordurkinn																		
4. Blágilsjökull																		
5. Hólatindajökull																		
III. EYJAFJALLAJÖKULL																		
1. Gígjökull																		
2. Seljavallajökull																		
IV. MYRDALSJÖKULL																		
1. Sólheimajökull W																		
» E																		
Jökulhöfud																		
V. VATNAJÖKULL																		
1. Skeidararjökull W																		
» E																		
2. Morsárjökull																		
3. Skaftafelljökull																		
4. Svinafellsjökull N																		
» S																		
5. Virkisjökull(Falljökull)																		
6. Gljúfursarjökull																		
7. Stígárjökull																		
8. Hólárjökull																		
9. Kvíárjökull																		
10. Breidamerkurjökull W																		
» E																		

FORANDRINGER VED ISLANSKE JOKLER 1930 - 1947 (OG 1904-1980) (suite)

	'30 /31	'31 /32	'32 /33	'33 /34	'34 /35	'35 /36	'36 /37	'37 /38	'38 /39	'39 /40	'40 /41	'41 /42	'42 /43	'43 /44	'44 /45	'45 /46	'46 /47
11. Fellsjökull						-96	-54	90	0	-125	-57	18	5	1	2	28	16
12. Brókarjökull						-13	1	12	27	11	34	10	7	7	10	10	23
13. Birmujökull						-18	10	5	60	50	31	24	10	2	8	20	30
14. Eyvindstungnakollur						-25	57	59	28	22	85	30	45	28	32	13	19
15. Heinabergsjökull S						-450	170	43	70	44	78	24	43	39	20	68	28
» N						-250	80	24	36	10	130	60	44	100	108	3	18
16. Fláajökull W						-400	108	42	75	67	100	34	42	49	49	46	25
» E						-550	40	55	9	44	91	135	142	98	32	16	31
17. Hoffellsjökull W							20	0	10	5	8	12	10	3	24	19	21
» E																	
18. Hoffellsdalsjökull						-60	10	30	40	-43	-7	-30	-20				
VI. HOFSJÖKULL																	
1. Blágnipujökull						-72		15		90		64		34			
2. Nauthagajökull						-73		152		10	26						
3. Múlaejökull						-140		167		36	55	6		12			
VII. LANGJÖKULL																	
1. Fúlakvísl						-41		40			-131					-50	
2. Hagafellsjökull vestri								-160			-170						
3. Hagafellsjökull systri								-500			-250						
VIII. HRÚTAFELL																	
1. Nordurkinn E						-50		20	25	56	20	25		50		45	
» M						-40		10	16	29	20	15		24		34	
» W						-10		35	38	8	50	15		60		40	
2. Nordvesturjökull						-65		18	25	25	30	20		46		45	
(NW-glacier)																	
IX. KERLINGARFJÖLL																	
1. Lodmundarjökull						-36		7		13	7	7		4		2	8

Reykjavík, 15.III 1948.

NORWAY

	1939 -	1940 -	1941 -	1942 -	1943 -	1944 -	1945 -	1946 -	1947
Bôyabre	— 50	— 90	— 15	—115	— 10	— 30	— 20		
St. Suphellebre	— 13	— 13	— 15	+ 20	— 5	— 32	— 17		
L. Suphellebre	— 30	— 70		(— 200)	15	inaccessible			
Austerdalsbre	— 38	— 86	—187	— 27	— 31	— 16	— 11	— 17	
Tunsbergdalsbre	— 19	— 5	— 5	— 16	— 18	— 37	— 17	— 43	
Bersetbre	— 32	— 29	— 66	— 60	— 42	← (—100)	→		
Nigardsbre	— 28	— 41	— 19	— 38	— 10	— 43	— 35	—113	
Fåbergstølsbre	— 33	— 33	— 17	— 53	+ 7	— 59	— 25	— 42	
Lodalsbre	— 22	— 22	— 35	— 32	— 15	— 28	— 65	— 33	
Stageholtbre	— 12	— 30	— 30	— 29	— 16	— 16	— 34	— 25	
Mjølkevoldsbre	— 23	—120	inaccessible						
Briksdalsbre	— 53	— 47	— 45	— 51	— 44	— 42	— 60	— 64	
Åbrekkebre	— 61	— 78	— 44	— 80	—113	— 97	—204	—105	
Bôdalsbre	— 20	— 10	— 13	—103	— 83	— 19	— 35	— 71	
Kjenndalsbre	— 74	—120	— 37	—174	—332		— 80	— 62	
Styggedalsbre	— 13	—9½	— 2	—3½	—5½	— 3	— 6	— 13	
Engabreen	—172	— 40	— 65	no reports received					
Fonndalsbre	—124	— 29	— 14	do.					

Buarbre and Bondhusbre have been inaccessible during whole the period.
K. Fagri,
Bergm.

	1935 -	1936 -	1937 -	1938 -	1939 -	1940
Buarbre	— 31	—129	— 45	— 62	— 43	
Bondhusbre	— 24½	— 28½	— 44	— 53½	—ca 200	
Bôyabre	— 35	— 45	— 50	— 20	— 50	
St. Suphellebre	— 3	— 35	— 13	— 47	— 13	
L. Suphellebre	— 43	— 40	— 15	— 90	— 30	
Austerdalsbre	— 36	— 23	— 11	— 36	— 38	
Tunsbergdalsbre	— 26	— 37½	— 10½	— 19½	— 19	
Bersetbre	— 39½	— 49	+ 4	— 41	— 32½	
Nigardsbre	+ 5	— 17	— 21½	— 49½	— 28	
Fåbergstølsbre	— 7½	— 75	— 18½	— 29	— 33	
Lodalsbre	— 23	— 16	+ 12	— 20½	— 22	
Stegaholtbre	— 4	— 9	— 1	— 39	— 12	
Mjølkevollsbre	— 39	—100	—101	—203	— 23	
Briksdalsbre	— 43	— 31	— 37	— 50	— 52	
Åbrekkebre	— 39½	— 25	— 43½	— 61½	— 61½	
Bôdalsbre	— 6	no data	— 25	— 44	— 20	
Kjenndalsbre	— 17½	— 47½	— 44½	— 24	— 74	
Styggedalsbre	— 4	— 12	— 5	— 20½	— 13	
Åmotbreen	— 46½	— 9	— 35	— 23	— ca100	
Engabreen	— 4	—150	— 43½	—481	—172	
Fonndalsbre	— 1	— 6	—120	— 55	—124	
Rembesdalsskåkje	— 10	— 12	— 4½	— 5½		
Nordfjordsbreen (Helgeland)	— 31	— 26	—186	—193	— 75	

Tillbakagangen av glaciärer i JOTENHEIMEN 1945 - 46
enl. mätningar av W. (och Nils) WERENSKIOLD.

Verslegjuvbre		
Heimre Illåbre	— 11.7	
S. Illåbre	— 33	
N. Illåbre	— 31.5	
Leirtjönbre	— 57.5	
Storgjuvbre	— 33.7	
Storbre i Leirdal	— 43.30	
Vestre Memurubre	— 95.2	(1938 — 43 150 m)
Svartdalsbre	— 8	
Slettemorkbre	— 35	
Langdalsbre	— 15.5	

RECENT GLACIOLOGICAL INVESTIGATIONS IN SWEDEN

by

Dr. H. W:SON AHLMANN, Professor, Stockholm

SUMMARY

The Swedish glaciers have been in recession during the last decades.

The Karsa Glacier (1,95 km²), situated on 68° 20' N and 18° 20' E, is the only one which has been measured since the beginning of this century. After surveying and introductory investigations in 1925-26 according to Ahlmann's principles (Ahlmann, London, 1948), the studies were continued by C. C. Wallén in 1942-46 by an enlarged programme.

The mean annual recession of the glacier was in 1909-39 3,4 m and in 1939-47 7,7 m. The lowest part — between 800 and 900 m above sea level — has decreased from 0,15 km² in 1925 to 0,04 km² in 1944.

The decrease in volume of the glacier has been estimated as 1,2 mill. m³ of water per km² per annum between 1926 and 1943.

The regime of the glacier appears from the following table (all figures in mill.m³ of water) :

Budget year	Accumulation	Ablation	Total regime	Net loss
1941-42	2,3	3,9	6,2 (= 3,1 pr km ²)	1,6 (0,8 pr km ²)
1942-43	3,7	4,1	7,8 (= 3,9 » »)	0,4 (0,2 » »)
1943-44	3,6	3,9	7,5 (= 3,3 » »)	0,3 (0,2 » »)
1945-46	3,5	—	—	—

The studies have mainly dealt with the relation between ablation and different meteorological factors. During all periods of investigations (involving four periods in the beginning of the ablation season, three during the middle of the summer and one in early autumn) the total insolation, ablation, temperature, humidity and wind have been recorded. Furthermore, observations series exist for the variation of temperature, humidity and wind in the air layer nearest the snow surface.

It has been stated that the variation with height of the different meteorological factors is much more complicated over a small glacier like the Karsa than over a plateau glacier like Isachsen Plateau in West Spitsbergen where the same problem has been previously studied by Sverdrup (Sverdrup 1935). The wind and temperature vary over the snow surface according to different laws, and the temperature gradient is dependent on the local weather conditions as well as on the character of the airmass to a much greater degree than was previously surmised. It is, therefore, necessary to adopt new interpolation formulas for the variations in temperature and wind, when calculating the transportation of heat to the snow or ice surface.

Preliminary studies have given the following results :

Time	Melted owing to radiation income	Evaporated owing to radiation income	Melted owing to convection	Melted owing to condensation
17.5-11.6 (snow)	61%	2%	33%	4%
20.6-10.7 (snow)	49%	0%	36%	15%
19.7-17.8 (snow)	30%	0%	48%	22%
19.7-17.8 (ice)	51%	0%	33%	16%

It should be emphasized that the figures for the radiation have probably to be lowered by a few percent, and those for the convection and condensation raised, but this will have no influence upon the general results. It is interesting to find that the evaporation is small.

We have also been able to show how the importance of convection and insolation changes with different types of air mass.

In general the convection must have the greatest importance for the ablation and thus for the recession of the glaciers during the last decades.

The results of these studies will be published in 1948 or 1949.

The Stor Glacier (3,3 km²), facing eastward in the Kebnekaise massif, situated on 67° 50' N and 18° 30' E has since 1945 been systematically investigated under the direction of V. Schytt, and in accordance to Ahlmann's principles (Ahlmann, London 1948).

The mean annual recession was in 1922-44, 8 m; in 1945-46, 17 m and in 1946-47, 22 m.

The decrease in volume between 1922 and 1946 is estimated as being 70 mill. m³ of water, or on an average 2,8 per year (= 0,8 per km²).

The accumulation amounted in the budget year 1945-46 to 3,5 mill. m³ of water, in 1946-47 to 3,2 mill. The latter figure is based on a great number of shafts and on more than 1000 soundings.

The regime of the glacier was (in mill. m³ of water) :

Budget year	Accumulation	Ablation	Total regime	Net loss
1945-46	3,5	5,5	9,0 (= 2,7 pr km ²)	2,0 (= 0,6 pr km ²)
1946-47	3,2	9,6	12,8 (= 3,9 » »)	6,4 (= 1,9 » »)

The rate of movement is determined by 30 stakes bored into the snow or ice. On the axis of the middle part of the glacier tongue the values was 14 m from August 1946 to August 1947; the winter figures here were 50 percent less than the summers'.

The air temperature was recorded on the uppermost part of the glacier (1480 m above sea level), on the side of its margin (1140 m) and at 680 m. About 200 ablation values, varying between 1 and 12 cm of water per 24 hours, from different parts of the glaciers will be correlated with the temperature on the respective parts, varying between 3° and 11° C.

The course of the temperature in the snow cover has been examined for determining the coefficient of heat conductivity, the recrystallization and redeposition of fluid water in the snow. Apart from melting occurring on the snow surface there is probably also an internal melting.

The Rabot Glacier (4,4 km²) facing westward in the same massif, had a decrease in volume between 1922 and 1946 of about the same magnitude as that of the Stor Glacier, but its annual regime is less.

The Kebnekaise massif is crested with a small, ridgeshaped glacier, forming the highest point in Sweden. This glacier has also diminished; its top from 2123 m above sea level in 1902 to 2114 m (or possibly less) in 1947.

The investigations in Kebnekaise are planned to be considerably enlarged in 1948.

Literature

- AHLMANN, H. W:SON u. TRYSELIUS, O. : Der Karsa-Gletscher (1925-26). Geografiska Annaler, Stockholm 1929.
- AHLMANN, H. W:SON u. LINDBLAD, Th. : Die Grössenveränderungen des Karsajökels 1909-29. Geogr. Annaler, Stockholm 1940.
- SCHYTT, V. : Glaciologiska arbeten i Kebnekajse, Ymer, Stockholm 1947.
- AHLMANN, H. W:SON : Glaciological Research on North-Atlantic Coasts. Monographs, I, Royal Geographical Society, London 1948.
- SVERDRUP, H. U. : The Ablation on Isachen's Plateau. Geogr. Annaler, Stockholm 1935.

GLACIER OBSERVATIONS IN THE CANADIAN CORDILLERA

by

V. MEEK,

Controller, Dominion Water and Power Bureau, Department of Mines and Resources

In connection with studies of the water resources of the mountainous rivers of British Columbia and Alberta, the Dominion Water and Power Bureau, Department of Mines and Resources, in 1945, initiated an annual survey of certain typical glaciers in the Canadian Cordillera. In addition to obtaining glacier information of general scientific value, the ultimate purpose in view is the determination of the effect of glacier variation on run-off, particularly the amount of run-off which may be expected in future from our glacier resources. A limited number of glaciers were selected for observation purposes, each reasonably accessible and considered to be typical of the general area of location. The studies in the Coast and Selkirk Ranges have been conducted under the direction of Mr. C. E. Webb, District Chief Engineer, Vancouver; those in the Rocky Mountain Range, under Mr. O. H. Hoover, District Chief Engineer, Calgary.

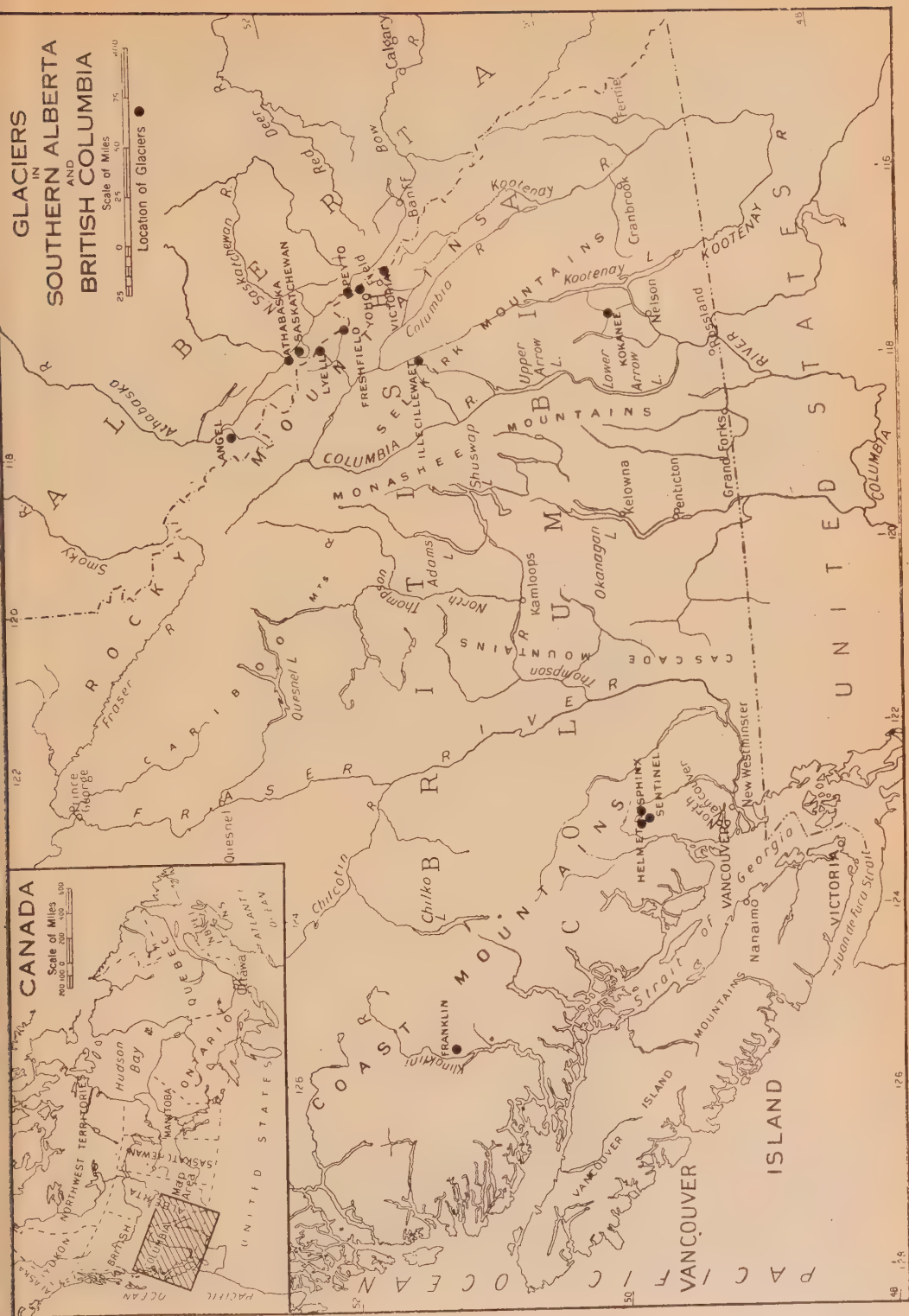
Glacier Phenomena

Glaciers are formed by great depths of snow accumulating in mountain basins at high altitude; the weight of the snow, assisted by surface melting, causes the lower layers to compact and to turn into ice. Under the pressure exerted, together with gravitational effect, the ice slowly extrudes through the valley outlets of the basin, the flow resembling that of a river. When the slowly moving mass of ice in the valley reaches lower altitudes, melting takes place during the summer months, forming glacial streams. When the annual amount of melting equals the rate of ice flow, the glacier becomes more or less stabilized, advancing in winter and retreating in summer; if the annual melting exceeds the forward movement, the toe of the glacier retreats from year to year. Our present glaciers are the remnants of the continental ice cap which once covered a large part of the northern half of the continent. « Alpine » glaciers are those flowing directly the upper icefield; « Piedmont » glaciers are formed by a union of one or more Alpine glaciers into a single stream.

The year to year changes in individual glaciers depend upon the amount of ice supplied from the upper icefield, the rate of ice flow and the quantity melted away on the tongue; if more ice is supplied than is melted, the glacier advances, while if melting exceeds supply, recession takes place. Annual temperature, precipitation, sunshine, amount of moraine covering and the topography of the valley modify the result so that many factors are involved. There may be also a considerable time lag before gradual changes in the icefield are reflected by the valley glacier. Long-term general variations of regional glaciers are caused by changes in climate, particularly the factors of temperature and precipitation. At the present time sufficient data have not been accumulated to determine the physical laws governing glacial phenomena.

Observations by Alpine Club of Canada

Although glacier observations in Canada under governmental direction began only in 1945, sporadic observations and studies of the variations of a number of glaciers in the mountainous regions of British Columbia and Alberta have been made by members of the Alpine club of Canada over a long period of years, the first on record being for the year 1887. In addition to photographs, the recorded observations usually have included the distance from a reference point to the nearest ice of the glacier toe; a limited number of observations also were made of the rate of glacier flow. Reports on these observations, together with excellent articles on the formation, characteristics and phenomena of glaciers, have been published in the Canadian Alpine Journal from time to time; three of the more noted investigators and writers were, — W. H. Sherzer, W. S. Vaux and A. O. Wheeler.



The observations of Canadian Glaciers recorded in the *Canadian Alpine Journal* reveal a general and practically continuous recession since 1887. This recession coincides with a similar glacier retreat in the Swiss Alps so that the earlier pattern of Canadian glacier variation probably was somewhat similar to that in Europe. In brief, beginning in 1812, there was a general advance of all glaciers which reached a maximum in 1825; a period of decrease then set in, followed by a less positive increase which reached a maximum about 1850; then ensued a period of marked decrease and in 1870, all the glaciers were definitely retreating. The period 1875 to 1894 was indecisive, certain glaciers advancing and others retreating; since then the decrease has become almost universal and has continued more or less positive in character until the present time.

Records of Glacier Variation

In 1894 an International Commission on Glaciers with headquarters at Lausanne, Switzerland, was appointed by the International Congress of Geologists meeting at Zurich and a general record of glacier variation throughout the world was kept for many years, reports being received from the Canadian Alpine Club covering observations in Canada. The Commission ceased to function during World War I but was reconstructed under the auspices of the Section of Scientific Hydrology of the International Union on Geodesy and Geophysics at its full meeting at Stockholm in 1930. Glacier data also have been compiled by the Smithsonian Institute at Washington. The studies underway by the Dominion Water and Power Bureau will provide current data on glacier conditions in Canada.

Methods of Observation

The methods of observation used to date by the Bureau include the following procedures :

1. The setting up of fixed reference marks near the toe of each glacier under observation. These markers are tied in to permanent topographical features.
2. The annual measurement of the distance from a reference point or base line to the nearest ice of the toe; or alternatively, the annual mapping of the forefoot. In the latter case, successive yearly plottings show clearly the average advance or recession across the whole forefoot.
3. The setting up of camera stations from which photographs can be taken annually to show overall changes in the glacier.
4. The study of the rate of glacier flow by means of plaques or markers set out on the ice along a base line crossing the glacier some distance above the toe. The base line is set perpendicular to the direction of flow and the deviation of the markers from the base line is measured trigonometrically each year. Field trials of different markers are under way.
5. Determination of annual ablation by means of a surface cross-section line run across the glacier tongue perpendicular to the line of flow and some distance up from the toe.
6. It has been recommended that aerial photographs be taken of each glacier at about five-year intervals to cover the general outline of the glacier and adjoining icefield. These would indicate the general shrinkage or extension of the supply reservoir which is actually more important from the long-term run-off viewpoint than the yearly local change of the glacier tongue.

Glaciers Under Observation

For the purposes of the Dominion Water and Power Bureau studies, typical glaciers in each mountain range were selected. In the Coast Range, two areas, Garibaldi Park and Mount Waddington, are comparatively accessible and considered to be representative of the range; the Helmet, Sentinel and Sphinx Glaciers were chosen in Garibaldi Park and The Franklin Glacier in the Mount Waddington area. In the Selkirks, Kokanee Glacier in the southern portion and the Illecillewaet in the more northerly region were adopted. In the Main Range of the Rockies, glaciers in the Banff and Jasper National Parks have been used including Victoria, Peyto, Saskatchewan, Athabaska, Freshfield and Lyell Glaciers; the Yoho and Angel Gla-

ciers were also inspected. The distribution of these glaciers and their specific locations are shown on the accompanying map which covers adjoining parts of British Columbia and Alberta; the general location is indicated on the inset map of Canada. A typical view of each glacier, selected from those taken in 1945, 1946 and 1947, is shown in the appended photographs.

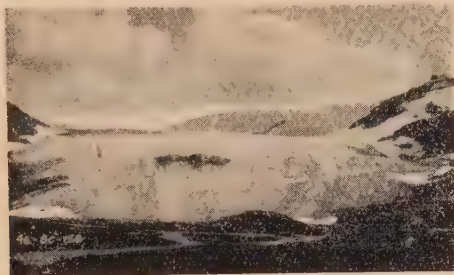
Records of the observations by the engineers of the Dominion Water and Power Bureau covering two full years are now available for most of the above glaciers. A short description of each glacier, together with a brief review of previous and current observations is given below. The observations are also summarized in tabular form.

COAST RANGE

1. *Sentinel Glacier*

The Sentinel, Sphinx and Helmet Glaciers are located in Garibaldi Park, about 60 miles north of Vancouver. The Sentinel is typical of all the glaciers in the Park. It is the Piedmont type being composed of several smaller glaciers which unite into a short trunk. Observations for the year 1935-36 showed a mean recession of 240 feet on two shallow projections of the toe. In 1945 the base line previously used could not be located but photographic evidence indicated a recession of about 900 feet from 1935 to 1945, being 90 feet per year. For the two-year period 1945-1947 the recession was 141 feet or 70 feet per year. The forward movement during 1946-47 was 10 feet.

Observer — P. W. Strilaeff, Field Engineer, Vancouver.

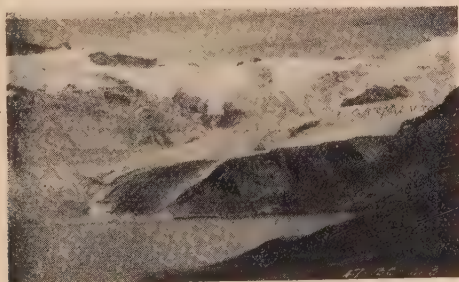


Nº 1

2. *Sphinx Glacier*

The Sphinx Glacier lies two miles north of the Sentinel. Its forefoot is very indefinite and irregular; it is difficult to approach at present owing to huge blocks of ice near the toe. The moraines deposited by its recession are undisturbed and provide an excellent geological record of its retreat. No measurements were made but photographs were taken in 1945, 1946 and 1947.

Observer — P. W. Strilaeff.



Nº 2

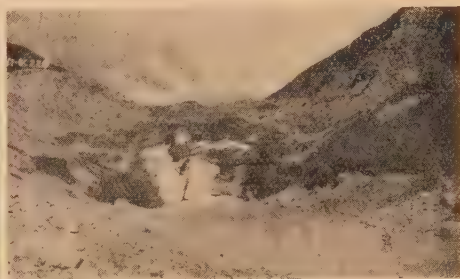
3. *Helmet Glacier*

Helmet Glacier is typical of the « Alpine » basin type of glacier with a northern exposure. It has two tongues about half a mile apart, on each of which observations were made in 1935 and 1936 by the British Columbia Mountaineering Club which showed a recession of 112 feet for the East Tongue and 85 feet for the West Tongue in one year. The observations of 1945-1947 show average annual recessions of 55 feet and 50 feet respectively. For the period 1935-1947 the West Tongue receded 546 feet or an average of 45.5 feet per year. The amount of forward movement on the East Tongue was 14 feet in the year 1946-1947.

Observer — P. W. Strilfaeff.



N° 3



N° 4

4. *Franklin Glacier*

This glacier is formed on the slopes of Mount Waddington which is situated about 165 miles northwest of Vancouver; it discharges into the Franklin River which flows into Knight Inlet near its head. Observations regarding its recession were instituted by Mr. D. Munday of the Canadian Alpine Club in 1927. During the period 1927-1945 the total recession was 4520 feet, being a yearly average of 251 feet. During the two-year period 1945-1947 the average was 292 feet.

Observer — A. V. Gallon, Field Engineer, Vancouver.

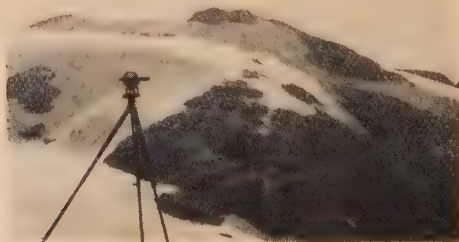
SELKIRK RANGE

5. *Kokanee Glacier*.

The Kokanee Glacier is located about 20 miles northwest of Nelson, on the north side of the highest ridge in the southern limits of the Selkirk Mountains. The glacier has two tongues about one mile apart, one at the head of Joker Creek and the other at the head of Coffee Creek, the tongues of both being in basins well below the main ice field.

Although no previous records have been compiled, it can be deduced from Geological Survey Map No. 203B, based on 1923 topographic surveys, that Coffee Creek tongue has receded about one-half mile and the Joker Creek Tongue about one-quarter mile since 1923. Observations covering 1945-1947 show the average annual recession of the Joker Creek tongue was 56 feet and that of Coffee Creek tongue 93 feet.

Observer — W. P. Harland, Field Engineer, Vancouver.



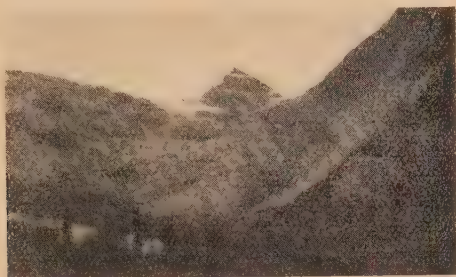
N° 5

6. *Illecillewaet Glacier*

This glacier which is located in the west side of the main divide of the Selkirks is three miles south of Glacier, on the Canadian Pacific Railway. It is the source of one of the main tributaries of the Columbia River. The glacier proper has an area of about nine square miles.

Records on the Illecillewaet date back to 1887 having been kept more or less continually since that date by members of the Alpine Club of Canada. Large alders growing near the toe in 1887 indicated that the glacier had been in this maximum position for many years. Since then the recession has been rapid but appears to be retarding since 1931. The average annual rate of retreat was 28 feet from 1898 to 1905, 83 feet from 1905 to 1911, 134 feet from 1911 to 1931 and 43 feet from 1931 to 1945. The average annual recession for the years 1945-1947 was 55 feet.

Observer — W. P. Harland.



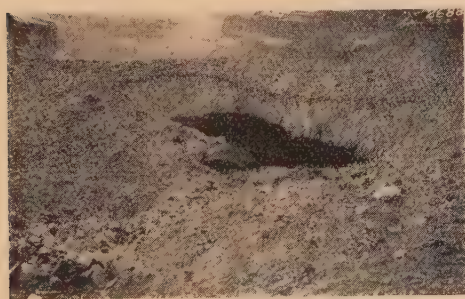
Nº 6

ROCKY MOUNTAIN RANGE

7. *Victoria Glacier*

The Victoria Glacier originates in Abbot Pass and is fed by snow and ice avalanching from the cliffs and hanging glaciers of Mounts Victoria and Lefroy. Its tongue lies in the valley of lake Louise at an altitude of about 6,000 feet, being buried under a mass of morainal material which makes observations of its true position very difficult. It has been under intermittent observation since 1898 the average rate of retreat for the period 1898-1912 being 18 feet per year; for 1945-1947 the rate was 33 feet.

Observer — W. T. McFarlane, Senior Engineer, Calgary.



Nº 7



Nº 8

8. *Peyto Glacier*

The Peyto glacier flows from the Wapta Icefield, the extent of which is about 40 square miles and general elevation 9,000 feet. It is the source of the Mistaya River. It was first photographed in 1897 and measurements were made in 1933 and at intervals thereafter. The average yearly recession since 1897 has been estimated at 42 feet, varying from 21 feet per year for the period 1897-1933 to 133 feet per year for the period 1939-1942. For the period 1945-1947 the recession averaged 45 feet per year and the rate of flow was 70 feet per year.

Observer — W. T. McFarlane.

9. *Freshfield Glacier*

This glacier belongs to the Freshfield Icefield which is situated on the Continental Divide between tributaries of the North Saskatchewan and Columbia Rivers, its drainage being principally to the Saskatchewan. The icefield has an area of about 22 square miles and the trunk glacier has a length of about nine miles discharging through a deep, narrow valley. It forms the headwater of the Howse River.

The glacier is easy of access, fairly smooth and ideal for purposes of record. Records, consisting either of photographs or measurements, are available dating back to 1902. The total recession since 1902 is 3,525 feet, an average of 80 feet per year for 45 years; during the last 25 years the recession averaged 104 feet per year. Observations by the Bureau were undertaken in 1947.

Observer — W. T. McFarlane.

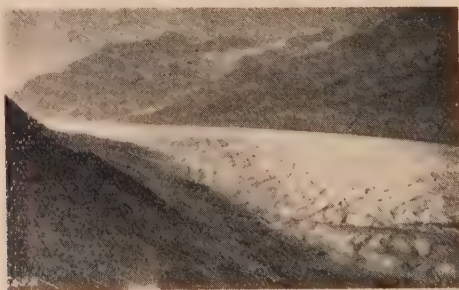


Nº 9

10. *Saskatchewan Glacier*

This glacier has its source in the Columbia Icefields and forms one of the headwaters of the North Saskatchewan River. The surface of the glacier is very rough and its toe is irregular. Surveys to record its recession began in 1945, no previous observations being known. During the two-year period 1945-1947 the total recession was about 250 feet.

Observer — W. T. McFarlane.

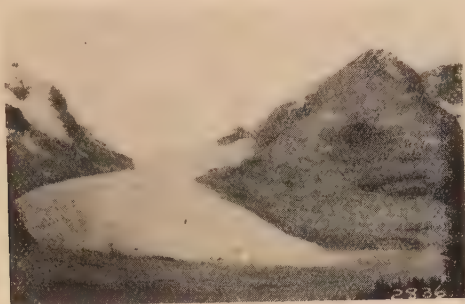


Nº 10

11. Athabaska Glacier

The Athabaska Glacier, also having its source in the Columbia Icefields, is considered to be excellent for observation purposes being fairly smooth and having a well defined toe. It is readily accessible from the Banff-Jasper Highway about 80 miles north of Lake Louise. Prior to the surveys covering both recession and flow which began in 1945, no records are known. The amount of recession for the two years was very irregular at different points of measurement varying from 50 to 220 feet. The rate of ice flow averaged 63.5 feet per year for the two years.

Observer — W. T. McFarlane.



N° 11

12. Angel Glacier

The Angel Glacier, which is at the base of Mount Edith Cavell in Jasper National Park, is more or less isolated and does not belong to any particular icefield. It was placed under observation in 1945 but is now not considered to be of value for record purposes as the tongue is not now connected to the upper glacier and the toe is completely covered with rock and difficult to locate. The amount of recession 1945 to 1947 varied from 40 to 200 feet.

Observer — W. T. McFarlane.



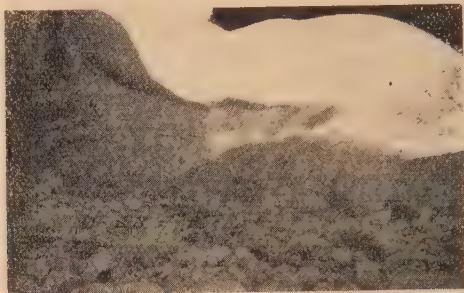
N° 12

13. Yoho Glacier

This glacier is the largest southern outflow from the Wapta Icefield and is located about 20 miles from Yoho Station on the Canadian Pacific Railway. It is the source of the Yoho River. When first discovered it presented a magnificent ice-fall which for many years attracted visitors. It was placed under observation by members of the Canadian Alpine Club in 1901 and intermittent records were kept until 1931. It was inspected in 1945 but was found to have retreated up the

valley until it is now a hanging glacier unsuitable for recording purposes; previous reference points could not be located. From the Alpine Club records the annual rates of recession were 1901-1904, 37 feet; 1906-1918, 33 feet; 1918-1931, 46 feet. The average rate of forward movement of the ice was 103 feet per year as observed for ten individual years between 1906 and 1918.

Observer — W. T. McFarlane.



N° 13

14. *Lyell Glacier*

This glacier forms part of the Lyell Icefield on the Continental divide and is approached from the Banff-Jasper Highway about 60 miles north of Lake Louise. It was inspected in September 1947, photographs taken and a reference base line established; however it is not regarded as well suited for observation purposes owing to its steep forefoot. It has apparently receded rapidly in recent years.

Observer — W. T. McFarlane.



N° 14

A weighted average for all the years of record prior to 1945 on ten of the above listed glaciers for which records are available for 1945-1947 shows a recession of 83.2 feet as compared with 88.6 feet for the period 1945-1947. This would indicate that the general rate of recession has accelerated during recent years. There are not sufficient data to draw conclusions in regard to variations in rate of flow.

Conclusion

It is generally agreed that the recession of our glaciers has been caused by a gradual long-term change in climatic conditions, primarily a slight increase in annual mean temperature; probably there was also been a lower rate of precipitation in the mountains and longer periods of sunshine. It is everywhere evident that, in former times, glaciers were of much greater extent than at present and that there has been a decrease and shrinkage for many years. Annual district variations in temperature, precipitation and sunshine determine if an individual glacier will advance or retreat over a short period of years and these same forces acting over longer periods of time determine the long-term trend. On individual glaciers the



Source of Confederation Glacier B.C. showing Mt Waddington in Background



Source of the Glacier White Mantle Range. — Knight Inlet B.C.

Summary of Observations

Glacier	Period of Record	Annual Rate of recession Feet	Annual Rate of Forward Movement Feet
COAST RANGE			
1. <i>Sentinel</i>	1935 - 1936	240	
	1936 - 1945	90	
	1945 - 1947	70	10 (1946-47)
3. <i>Helmet</i> — West Tongue	1935 - 1936	85	
	1936 - 1945	36	
	1945 - 1947	50	
East Tongue	1935 - 1936	112	
	1945 - 1947	55	14 (1946-47)
4. <i>Franklin</i>	1927 - 1934	221	
	1934 - 1945	270	
	1945 - 1947	292	
SELKIRK RANGE			
5. <i>Kokanee</i> — Coffee Creek	1923 - 1945	120	
	1945 - 1947	93	
— Joker Creek	1923 - 1945	60	
	1945 - 1947	56	
6. <i>Illecillewaet</i>	1898 - 1905	28	
	1905 - 1911	83	
	1911 - 1931	134	
	1931 - 1945	43	
	1945 - 1947	55	
ROCKY MOUNTAIN RANGE			
7. <i>Victoria</i>	1898 - 1903	17	147 (1900)
	1907 - 1912	19	
	1945 - 1947	33	159
8. <i>Peyto</i>	1897 - 1933	21	
	1933 - 1936	79	
	1936 - 1939	117	
	1939 - 1942	133	
	1942 - 1945	77	
	1945 - 1947	45	
9. <i>Freshfield</i>	1902 - 1922	95	
	1922 - 1937	87	71
	1937 - 1947	137	71
10. <i>Saskatchewan</i>	1945 - 1947	125	
11. <i>Athabaska</i>	1945 - 1947	135	63
12. <i>Angel</i>	1945 - 1947	120	
13. <i>Yoho</i>	1901 - 1904	37	
	1906 - 1918	33	103
	1918 - 1931	46	

rate of change may be affected by conditions of the valley such as steepness of the slope, width of tongue and amount of moraine covering. It is expected that the systematic observations now being made annually by the Bureau will yield data which will be invaluable for studies co-relating weather and climatic conditions with glacial phenomena.

☐ As glacial run-off constitutes an important part of the flow of many of our rivers, the effect of glacial recession on stream flow is of primary importance to the Bureau.

The measurement of glacial streams in close proximity to the glaciers of origin is usually not practicable owing to poor channel conditions in moraine material and would be of little value owing to the extremely capricious flow. On the lower reaches of stream largely of glacial origin more satisfactory records can be secured and these may be compared with those obtained on non-glacial streams. When sufficient data have been accumulated, studies will be undertaken in respect to the effect of glacial recession on stream flow, although these will be complicated by the many factors involved.

Description of Photographs

Glacier	Photographer	Date	Description
1. Sentinel	P.W. Strilaeff	Aug. 27, 1946	From a distance
2. Sphinx	»	Aug. 27, 1946	» »
3. Helmet	»	Aug. 28, 1946	Right side of west tongue
4. Franklin	A.V. Gallon	Sept. 12, 1947	From « C » marker
5. Kokanee	W.P. Harland	Sept. 9, 1945	From « A » reference point Joker Creek tongue on right.
6. Illecillewaet	»	Sept. 19, 1945	From reference point « W »
7. Victoria	W.T. McFarlane	Aug. 16, 1946	Toe showing moraine covering
8. Peyto	»	Aug. 19, 1946	From reference point
9. Freshfield	»	Sept. 12, 1947	From edge of timber
10. Saskatchewan	»	Aug. 23, 1946	From mountain side
11. Athabaska	»	Aug. 26, 1946	From Wilcox Mountain
12. Angel	»	Aug. 26, 1946	Toe from below
13. Yoho	»	Aug. 20, 1945	Toe from below
14. Lyell	»	Sept. 11, 1947	General view

GLACIAL FEATURES OF THE LLOYD GEORGE MOUNTAINS, BRITISH COLUMBIA

by

N. E. ODELL

The Lloyd George Mountains, situated in approx. N. lat. 58, W. long. 125, and previously unexplored, were investigated during the summer of 1947 by an Anglo-Canadian-American airborne expedition, which was sponsored by Mr. F.S. Smythe. Flying some 400 km. north from Fort St. James to a glacier-fed lake (Haworth Lake) at an altitude of 1160 m., the latter was used as a base to explore the range on foot from the southward. The range is a part of the Rocky Mountains proper, lying east of the Rocky Mountains Trench, and it is carved out of a great thickness of Palaeozoic limestones, which display tilted faultblock structure.

The dominating physiographic feature, from which rise the culminating peaks at c. 3000 m, all of which were ascended, is the George Lloyd Icefield, and its main névé (altitude 2500 m) covers some 130 sq. km; but with dependent glaciers its total extent must be about 260 sq. km. These outlet glaciers descend often by steep ice-falls to deep valleys lying at 1200 m, and they show evidence of slow recession from former terminal moraines not more than 1 km. distant.

The present summer snow-line lies at approx. 2400 m, and a rich alpine flora extends from heavily forested valleys to an upper limit of at least 2150 m.

A particularly interesting discovery, lying in a deep canyon-like valley, below the West Peak of the Lloyd George Group, was a large dying glacier entirely blanketed in its own moraine, upon which was a liberal spread of plants and small conifers up to 4.5 m in height. This stagnant glacier, some 4.5 km. in length, appears to have negligible nourishment either from a small cliff glacier at its head, or from avalanche snow from the precipitous rock walls enclosing it. The dead black ice could be seen in occasional exposures in sink-holes and at the borders of the glacier, and it appeared to extend (in depth) as far as the original terminal moraine at an altitude of about 1500 m. As far as is known, this glacier is the only one of its kind in the Canadian Rockies proper, in which the whole body of ice lies dead and stagnating beneath its mantle of moraine.

As to morphological relics of the former Cordilleran ice-sheet, no direct evidence was seen beyond erratic blocks in some valleys, the certain provenance of which was unknown. Farther south in the Rockies transported material has been found on some mountains up to at least 2450m.; but it is generally supposed that there has been land-elevation since the Glacial Epoch of over 200 m. The high peaks of the Lloyd George Mountains, and of the contiguous ranges, have, however, in many instances long stood above an old accordant summit level, and the glacial relics now seen are those of erosion and destruction.

The contiguous ranges to the north-west, dominated by Mount Roosevelt, Churchill Peak, etc. of c. 3000 m., which were examined from the air, showed a maturely dissected alpine topography with valley glaciers in some cases up to 6 or 7 km. in length.

6) THE VARIATIONS OF ALASKAN GLACIERS 1935-1947

by

FIELD

Since the last meeting of the International Commission of Snow and Glaciers in 1939, detailed studies have been made of the terminal portions of a number of Alaskan glaciers and a great deal of aerial photography has been undertaken for mapping purposes which provides data of great interest to the glaciologist. The general behavior of the glaciers of this part of the continent is fairly well-known, especially in the coastal areas, where observations are more easily made and, therefore, more frequent. With the exception of a few glaciers whose terminal advance or recession are measured every few years, data on the behavior of the glaciers can be given only in general terms based on a comparison of photographs taken at different times, presence or absence of terminal and marginal barren zones and the characteristics of the vegetation growing near the present ice limits. It is to be regretted that prior to this year, no serious glaciologic studies have been undertaken near the firn line or in the areas of accumulation of these glaciers, so that little is known of their regimens. However, studies are now in progress designed to obtain such information during next few years.

Since the latest information on Alaskan glaciers included in the 1939 report dated mostly from observations made in 1935, this report will consider the period from that year to the present. The order to be followed in discussing the various glaciers or groups of glaciers will be in geographical order proceeding along the coast from southeast to northwest and west; then east to west along the interior ranges.

Little specific data are available on the glaciers of the Coast Range between Portland Canal and the Stikine River. However, a thinning in the terminal portions accompanied by slight recession of the termini seems to predominate.

A number of the glaciers on the west slope of the Coast Range between the Stikine and Taku Rivers have been more carefully studied. Le Conte Glacier has receded about three-quarters of a kilometer in the last two decades and the lower part of the glacier has been thinned. The small glaciers of the fiord have also shrunk appreciably. The two glaciers at the head of Muddy River have also been receding. Patterson Glacier has receded nearly one kilometer since its maximum about 1900 when it was in contact with mature forest. From 1941 to 1946 the whole lower part of the glacier thinned considerably and the terminus receded about 200-300 meters. Baird Glacier has not receded more than 100 meters from the terminal moraine formed at its recent maximum when mature trees were overturned as late as 1935, but the surface of the whole lower end of the glacier has been lowered appreciably.

The glaciers of Holkham Bay have been visited a number of times. The terminus of Dawes Glacier after remaining almost stationary from 1923 to 1941, receded about 400 meters on the south side from 1941 to 1946 with little observable change on the north side. Vertical thinning of the lower glacier's surface during this period is estimated at about 80 meters. North Dawes Glacier maintained the rapid recession which has been going on since about 1922, amounting to an additional several hundred meters from 1941 to 1946. The South Brown Glacier, being a remnant ice mass cut off from its former source of supply, has continued its slow shrinkage beneath an ever increasing mantle of ablation moraine. Brown Glacier has shrunk in volume, but its terminus has not receded appreciably in the last decade. Sumdum Glacier continued its slow recession from 1935 to 1946. Like that of Dawes Glacier, the terminus of South Sawyer Glacier, after remaining virtually stationary from 1923 to 1941, receded about 200 meters on the north side and 500-600 meters on the south. Sawyer Glacier receded about 300 meters from 1935 to 1941 and 100-150 meters more from 1941 to 1946.

The rapid recession of North Dawes, South Brown and Brown Glaciers appear to be due in large part to loss of ice spilling over from the extensive icefields to the east at the head of Dawes and South Sawyer Glaciers. This suggests that the level of this névé has been appreciably lowered, perhaps in the order of 30 to 60 meters in the last century.

The glacier at the head of Farragut River has receded about one-half kilometer

since 1929 and has suffered tremendous vertical shrinkage. Here again, the source of supply has been virtually cut off, probably due to the lowering of a névé surface, and this glacier has now become a stagnating remnant. Speel Glacier appears to have been slowly receding and shrinking in recent decades.

The glaciers of Taku Inlet afford an interesting example of varied behavior. The terminal portion of Wright Glacier has thinned appreciably since 1929 with a recession of the west side of the terminus of at least 400 meters. However, the most advanced part of the terminus is still within a few hundred meters of mature forest. Nearby summit glaciers have also shrunk noticeably in the last two decades. Tal-sekwe Glacier appears to have receded only a few hundred meters in recent decades. Vertical shrinkage, however, has been considerable and the former lower tributaries from the west have all but disappeared since 1910. The terminal portion of both East and West Twin Glaciers shrank between 1941 and 1946 with East Twin showing a recession of perhaps 200 meters and West Twin experiencing virtually no change. Recession of these two glaciers has apparently been continuous since they were joined in a lobate terminus occupying the present site of Twin Glacier Lake as late as the 1890's.

Taku and Hole-in-Wall Glacier should be considered together as the latter is merely a distributary tongue of the former. Rapid advance of both has continued. Taku Glacier has advanced about 5 kilometers since 1900, much of which has been over ground previously occupied by mature trees. From 1937 to 1941, the advance varied from about 200 meters on the west side to a maximum of about 800 meters in the center. A further 400 meter advance occurred from 1941 to 1946 with a corresponding thickening of 100 meters or more throughout the terminal portion. The terminus now rests in comparably shallow water or is fronted by an extensive push moraine. The terminus of Hole-in-Wall Glacier, advanced some 200-300 meters between 1941 and 1946 and reached the flats bordering the Taku River.

Norris Glacier, whose terminus is less than $1\frac{1}{2}$ kilometers from the Taku has been receding slowly since a maximum about 1915 when the glacier was in contact with mature forest.

In the Lynn Canal area, Mendenhall Glacier has continued its steady recession ranging from 200 meters on the east side to 850 meters in the center between 1931 and 1945. Herbert and Eagle Glaciers have also experienced recession of several hundred meters and terminal shrinkage during the past two decades. Most of the glaciers of the Takhin and Tsirku Valleys have not experienced appreciable recession since 1910, but many show evidence of slight terminal thinning. As an exception, Tsirku Glacier has experienced a net advance of some 300-400 meter since 1910. This appears to have been contributed primarily by the two lowest tributaries on the north which have noticeably increased in volume. An unnamed glacier immediately east of the Tsirku has also advanced. Garrison Glacier has been receding slowly, but its terminus is still within a few hundred meters of mature forest. Rainbow Glacier also shows evidence of slow recession. The lobate terminus of Davidson Glacier apparently receded slowly from 1931 to 1945 accompanied by some thinning of the terminal area. Other glaciers of the Lynn Canal area have not been specifically examined, but appear to have been slowly receding.

Glacier Bay, whose glaciers have experienced tremendous net recession and shrinkage amounting to a maximum of over 100 kilometers since the 18th Century, now presents a more complex picture. Recession and shrinkage has continued in all the glaciers of Muir Inlet. Muir Glacier, itself, receded about $10\frac{1}{2}$ kilometers from 1935 to 1947 with a lowering of the ice surface at the latter position of some 300 meters. The inlet has been increased in area some 15.5 square kilometers in this 13 year period. McBride Glacier, a former tributary, became a separate glacier about 1944; Plateau Glacier receded about 3 kilometers from 1935 to 1946; while Casement and Morse Glaciers have continued their slow recession. The glacier in Adams Inlet which was about 180 meters thick and 65 square kilometers in extent in 1931, had practically disappeared by 1945 and its former tributary glaciers, Girdled and Adams, were receding up their respective valleys.

Carroll Glacier in Queen Inlet was receding and shrinking in 1935 and 1941, but in 1943 its lower portion was appreciably thickened and a minor advance apparently set in, which, however, had ended by 1946.

Rendu Glacier experienced a minor advance which culminated in 1935 or 1936 and has receded several hundred meters since that time.

In Tarr Inlet, Grand Pacific Glacier advanced about 760 meters from 1936 to 1947. The center and west side of the terminus now rest over the International

Boundary in Alaskan territory. The terminus of Margerie Glacier has remained in almost the same position since 1912, experiencing only minor oscillations during that time. In 1946, it was in a relatively advanced position. Some of the tributary glaciers of the Grand Pacific and the small hanging glaciers in Tarr Inlet also appeared slightly forward in 1946 and 1947 relative to the previous decade.

The glaciers of Reid Inlet (also referred to in some reports as Johns Hopkins Fiord or Inlet) have exhibited rather remarkable oscillations in the past two decades. After its recession of some 11 kilometers between 1912 and 1926, John Hopkins Glacier has readvanced. From 1935 to 1941, this ranged from about 1400 meters on the west margin to about 75 meters in the middle of the terminus. From 1941 to 1946, this slow advance continued and the whole lower part of the glacier increased in volume.

Five of the hanging glaciers of the inlet show an interesting similarity of behavior, namely recession from 1935 to 1941, followed by readvance to 1947. Toyatte Glacier reached tidewater in 1935, but terminated some 250 meters above tide in 1940 and 1941. The terminus of Kashoto Glacier in 1935 was at an elevation of about 120 meters and back to about 300 meters in 1940, then down again to around 120 meters in 1941 and at tidewater in 1946 and 1947. The lower part of Gilman Glacier shrank in volume from 1935 to 1941; the terminus was reported advanced in 1945, but back to its former position in 1946. In the latter year, its volume was greater than in 1941, and about that of 1935. Clark and Tyeen Glaciers both receded from 1935 to 1941 and readvanced between 1941 and 1946.

Hoonah Glacier appears to have increased in volume in its terminal area from 1935 to 1941, followed by no appreciable net change from 1941 to 1946, but some advance from 1946 to 1947. The terminus of Topeka Glacier shows no appreciable net change from 1935 to 1947, but a hanging glacier above it, advanced between 1935 and 1940 and again from 1940 to 1941.

The terminus of Lamplugh Glacier after showing little change from 1907 to 1937, receded approximately 760 meters by 1941, then readvanced by 1943 and again by 1945, followed by recession to 1946. The latter position was 100 meters or so ahead of that in 1941.

Reid Glacier about 1500 meters from 1935 to 1941, then more slowly from 1941 to 1946. Hugh Miller Glacier seems to have experienced continuous recession from 1935 to 1946 amounting to between 900-1500 meters. Maynard and Charpentier Glaciers receded several hundred meters from 1935 to 1941 and Geikie Glacier receded some 650 meters during that interval.

The glaciers of Glacier Bay thus show behavior since 1935 which may be summarized as follows: the glaciers of Muir Inlet, Geikie Inlet, Hugh Miller Inlet and Reid Glacier have apparently receded continuously during this time; Carroll and Rendu Glaciers have had minor advances within a pattern of net shrinkage; Grand Pacific Glacier and Johns Hopkins Glacier have had a net advance; Margerie and Lamplugh Glaciers have oscillated within limits of about 1 kilometer; and several of the hanging glaciers, both in Tarr Inlet and Reid Inlet have experienced from minor to major oscillations. In general, the period from 1935 to 1941 was one of recession in the small glaciers, and 1941 to 1946, a period of readvance. Noteworthy perhaps, is the fact that the two big glaciers which are advancing have their sources in the High Fairweather Range, whereas the glaciers have receded more or less steadily, are in lower subsidiary ranges to the east of the main range.

No detailed studies have been made of the glaciers along the west coast of the Fairweather and Brabazon Ranges between Cape Spencer and Yakutat Bay in recent years. Aerial photographs, however, indicate that the La Perouse Glacier has undergone no very great change between 1934 and 1947, though slight marginal withdrawal is evident. Its ice is apparently still within 100-200 meters of the mature forest invaded in the advance which culminated about 1910 or 1911. A large unnamed glacier to the southwest, however, shows more terminal shrinking and recession between 1934 and 1947, but still remains within a few hundred meters of the forest.

In Lituya Bay, the Lituya and Crillon Glaciers were at their maximum positions, in contact with mature trees, in 1937, but as far as known, have not been observed since. In the Alsek Valley, aerial photographs indicate recession of the Alsek and Melbern Glaciers with large areas of their terminal portions reduced to a stagnant condition. To the west, in the Brabazon Range, the various westward flowing glaciers glimpsed from the air give the impression of steady, slow recession.

In Yakutat Bay, and Russell Fiord at its head, a series of ground and aerial obser-

uations were made in 1946 and 1847. These reveal that Fourth Glacier has receded about 800 meters on the east, and 1600 on the west since 1909, while Hidden Glacier has receded nearly 5 kilometers since 1909 and is now about 1600 meters back of its 1905 position. Nunatak Glacier receded about 10 1/2 kilometers from 1909 to 1946 accompanied by a maximum vertical shrinkage of at least 600 meters. Of this, about 1200 meters occurred between 1934 and 1938 and about 800 meters from 1938 to 1946. A former tributary on the south has receded about 1 1/2 kilometers in the last decade. The terminal portion of Cascading Glacier has thinned and receded 90-120 meters since 1900, while a number of other small hanging glaciers both in Nunatak and Russel Fiords show no appreciable net change in the last three decades. In Disenchantment Bay, Variegated and Orange Glaciers have apparently experienced no appreciable net recession since 1909, but there may have been some vertical shrinkage. On the other hand, the terminus of Hubbard Glacier in 1946 was about 400 meters in advance of its position in 1910 and Turner Glacier showed slight advance and lateral expansion relative to its position in 1909.

No detailed observations of Malaspina Glacier have been made for half a century, but aerial photos suggest gradual withdrawal of its terminus and lowering of its surface. The same may be said for the Agassiz lobe. At the head of Icy Bay, the combined termini of Guyot and Tyndall Glaciers have receded nearly 24 kilometers since 1904, about 11 of which has taken place since 1913 and about 800 meters since 1934. West of Icy Bay, the termini of both Beare and White River Glaciers do not appear to have receded very rapidly in recent years.

Across the St. Elias Range on its eastern drainage, Wolf Creek Glacier in 1941 was found to be shrinking in its terminal area, but a number of its small tributaries were apparently advancing.

Bering Glacier is now receding slowly from terminal moraines formed during maximums which date about 100 years ago on the eastern and 50 years on the west. Barren zones on nunataks and along the margins also attest to gradual shrinkage of the volume of the piedmont area.

So far as known, the glaciers of the lower Copper River have not been studied in the last decade. However, aerial photographs suggest slow recession of the termini, shrinking of the terminal portions of the glaciers, and continued reduction of the stagnant ice masses remaining from earlier advances dating from perhaps a century or two ago. Sherman, Sheridan and Scott Glaciers near Cordova also appear to be shrinking slowly.

In Prince William Sound, a few specific observations are available. Valdez Glacier is apparently continuing its steady recession with an additional 30-60 meters from 1935 to 1941. Shoup Glacier's terminus remains virtually stationary, but there appears to have been some shrinkage from 1935 to 1943. Columbia Glacier which was experiencing a minor advance in 1935 had receded considerable by 1941. On the west end of the terminus, recession continued from 1941 to 1946, but on Heather Island there may have been an advance. Nevertheless, at this point the 1947 position appears to be in the order of about 300 meters back of the 1935 position, and the eastern terminus in 1947 was 100 meters or more back of its 1935 position.

Meares Glacier advanced into mature forest as much as 100-200 meters between 1935 and 1947. Yale Glacier has receded slightly on its west side and 100 to 200 meters on the east side during this interval. Harvard Glacier, however, has continued its advance with an additional 150 to 180 meters, most of which seems to have occurred since 1941. The hanging glaciers of College Fiord show behavior. The ice tongues on the southeast side draining from relatively low névés appear to be all in process of slow shrinkage and recession. The big hanging glaciers which reach tidewater on the northwest side of the inlet and have their source in high névés have behaved differently in this period. The terminus of Smith Glacier has not changed very much, although some advance may have occurred, but the terminal portion of the ice stream definitely appears to be thicker. Bryn Mawr Glacier seems to have changed very little from 1935 to 1941 and then to have advanced some 450 meters by 1947 to reform a lobate terminus occupying the small inlet formerly leading from the fiord to the face of the glacier. This lobate terminus was last in existence about 1914, but this recent advance seems to have carried the ice beyond the terminal moraine dating from that time and to be more advanced than for perhaps a century or more. Vassar and Wellesley Glaciers show little change although the terminal portion of the latter may have expanded very slightly during this period.

In Harriman Fiord, no appreciable change seems to have occurred in the termini of Barry, Cascade, Serpentine, Baker and Surprise Glaciers, from 1935 to 1947, but Coxe Glacier appears to have experienced a small net advance. The terminus of Cataract Glacier receded from tidewater to an elevation of 30-60 meters between 1935 and 1941 and has not undergone any significant net change since then. Harriman Glacier has continued its slow advance measurable at the west end of the terminus as about 75-90 meters. Close by, the much smaller Dirty Glacier, flowing from low névés, appears to have receded 75-90 meters during this same interval. Recession of Toboggan Glacier has been of the same order.

In Blackstone Bay, comparative photographs indicate that Marquette, Lawrence and Ripon Glaciers all show some terminal shrinkage and slight recession between 1935 and 1947, but no very rapid changes are observable on any of the glaciers.

Comparatively, few of the remaining coastal glaciers and those of the interior valleys have been observed in the last decades. Much information of record is to be had from the aerial photographs taken during the last decade, but these have not yet been correlated. There are two exceptions, however, one of which, Black Rapids Glacier, in the Alaska Range excited much interest in 1936 and 1937, when an advance of about 4.8 kilometers in 4 months was reported. Four years later, in 1941 no appreciable recession had occurred, but narrow marginal barren zones indicated that the terminal portion of the glacier was in the process of shrinking.

The second exception is the Yanert Glacier which rises also in the Alaska Range not far from the head of the Black Rapids and flows in the opposite direction. Aerial photos show that a significant advance was under way in both 1941 and 1943. However, no data on its subsequent behavior is available.

Nearly all the remaining glaciers of interior Alaska have conspicuous terminal and marginal barren zones indicating slow to rapid recession in their terminal portions in recent decades. Advancing glaciers in the interior ranges are more of an exception than along the coast. Some areas of the interior appear to be in a stage of deglaciation approaching in intensity that of the Muir Inlet basin, but observations are as yet extremely inadequate.

It is of interest that during the period 1935 to 1947, the Baird, Taku, Crillon and Lituya Glaciers of Southeastern Alaska have attained positions farther forward than at any time at least since the 18th Century and probably, longer, and that the Meares, Harvard, Harriman and possibly the Bryn Mawr in Prince William Sound have done likewise. Near these glaciers other termini have receded only within the last few decades from comparable maximum positions, and many others, though not now at maximum positions, remain within a few hundred meters of mature trees. A cursory analysis of existing aerial photographs covering almost all Alaskan glaciers, strongly suggests that the majority of glaciers which are now receding are withdrawing from terminal moraines formed during maximum advances in the 18th and 19th Centuries. Detailed studies are needed of selected representative terminal moraines to establish their approximate age. However, if they do date as suggested from the last two centuries, the present maximums being established along the coast may belong to the last phase of this period of glacial expansion which has had its counterpart in many other parts of the world.

The aerial survey of the northwest part of the continent has made it possible to ascertain the presence and extent of glaciers either not known to exist because of their small size or their position in remote unmapped areas. Data on small cirque glaciers east of the Coast Range in British Columbia and Yukon Territory is now available while the considerable glaciers of the Brooks Range in the north, the Kilbuck Mountains in western Alaska, and the Aleutian Range in the southwest are revealed in much greater detail than previously. The small and large glaciers of much of Northwestern Canada have similarly been exposed to view for the first, time in a way that makes possible a clearer understanding of their characteristics the moraine patterns and the relationship between the areas of accumulation and ablation.

The meteorological records for this period 1935 to 1946 indicate that summer temperatures were higher than normal in the areas where glaciers exist in Alaska and though total annual precipitation increased at 13 out of 14 selected stations, the amount of annual snowfall decreased on an average at 11 of these same 14 stations. Departures from the mean temperatures varied greatly in different places but in general, this period was one of relatively great summer ablation and deficient snow accumulation compared to conditions prevailing during the last few decades. The glaciers therefore, may be considered to have suffered appreciably both by greater melting,

and lessened snow supply. In other words, conditions affecting the regimen of glaciers were negative to a greater degree than previously. The general result of widespread, slow, steady recession is in accordance with the trend of the meteorological data. The advancing glaciers may be reacting to special conditions or to the fact that their sources are at higher elevations, which may be receiving a greater accumulation of snow due to a regional elevation of the zone of maximum snowfall. At present, there is insufficient data from which to determine this from direct observations. The only evidence at hand is the behavior of the termini, the meteorological records at the different stations, all of which are at elevations corresponding to the termini of the nearest glaciers or below them, and the general topographic characteristics of the glaciers themselves.

An attempt to determine whether conditions favored a relatively high or low firn line on the glaciers in the different years has been made from the records at fourteen selected meteorological stations. The factors taken into account are the temperature of the ablation season and the amount of precipitation occurring since the end of the previous ablation season. The results are given in the following table and are to be considered purely as indicating the tendency in this respect among the glaciers affected by the particular set of meteorological conditions at these stations.

Relative Position of firn line

Based on the departure from the mean of the temperatures of the ablation season and the precipitation during the preceding period from October to April.

Year	1935	1936	1937	1938	1939	1940	1941	1942	1943	1944	1945	1946
Southeastern												
Alaska												
Ketchikan	—	HH	H	H	—	HH	HH	HH	—	H	L	—
Sitka	—	H	L	—	L	H	HH	H	—	L	L	—
Wrangell	X	HH	H	H	LL	—	HH	H	L	—	L	—
Juneau	L	H	—	—	L	H	H	H	—	H	L	—
Annex Creek	—	H	—	L	L	—	H	H	—	L	L	—
Haines	—	H	—	L	L	—	H	H	—	L	L	—
Yakutat	X	X	H	—	L	X	—	H	X	—	—	X
South Central												
Alaska												
Kennecott	—	H	L	H	L	H	—	—	H	—	L	HH
Valdez	X	HH	—	—	L	H	—	H	X	L	L	X
Seward	—	HH	H	X	X	—	—	H	H	—	—	H
Anchorage	—	H	—	H	—	H	—	HH	—	—	—	—
Matanuska	X	H	L	—	—	H	H	HH	H	—	L	L
Talkeetna	L	H	L	—	L	H	H	H	H	—	—	—
Central Alaska												
Fairbanks	L	—	LL	—	—	—	—	H	H	—	—	—

Key: X — No record.

— Firn line apparently near mean position.

L — Firn line relatively low compared to the mean position.

LL — Firn line very low compared to the mean position.

H — Firn line relatively high compared to the mean position.

HH — Firn line very high compared to the mean position.

Note: LL and HH represent conditions where both temperatures and precipitation favor a low or high firn line so that the combination of both factors tends to produce a very considerable departure in the position of the firn line from its mean position.

by

Dr. HANS J. SPANN

Since June 1945, nine bench marks, (consisting of a prominent cairn) have been established in the Southern and Central parts of the Peruvian Andes in order to determine the actual position of glacier termini and to measure possible glacier recession by later surveys. Those bench marks are distributed from S to N as follows :

1. — « San Francisco Glacier » ($14^{\circ} 37'$ South $69^{\circ} 20'$ West) at the « Ananea Mountain » situated in the « Cordillera de Carabaya » which is a continuation of the « Cordillera Real » coming from Bolivia, and which forms the hydrographic divide between the Titicaca- and Amazon Basins. Altitude of bench mark approx. 14604 ft. (see : Bol. del Inst. Geol. del Perú, No. 5, Lima 1946, Phot. 1 y 4).

2. — « Ampato Glacier » at the eastern slopes of « Volcano Ampato » ($13^{\circ} 47'$ South $71^{\circ} 14'$ West) North of the City of Arequipa. Altitude of bench mark approx. 18077 ft. (see : Bol. del Inst. Geol. del Perú, No. 5, Lima, Phot. 22, 23, 24).

3. — « Paccopampa Glacier » ($13^{\circ} 26'$ South $71^{\circ} 10'$ West) at the Southern slopes of « Peak Paccopampa », belonging to the « Ausangate Range » which forms a continuation of the « Cordillera de Carabaya » to the northwest. Altitude of bench mark approx. 14764 ft. (see : Bol. del Inst. Geol. del Perú, No. 5, Lima 1946, Phot. 19, 20).

4. — « Incachiriasca Glacier » ($13^{\circ} 16'$ South $72^{\circ} 36'$ West), at the southern slopes of « Mount Salcantay », enormous peak northwest of the City of Cuzco between the Apurimac- and Urubamba Valleys. Altitude of bench mark approx. 15076 ft. (see : Bol. del Inst. Geol. del Perú, No. 5, Lima 1946, Phot. 13).

5. — « Umantay Glacier » ($13^{\circ} 15'$ South $72^{\circ} 35'$ West) at the WNW slopes of « Mount Umantay » north west of the City of Cuzco between Apurimac and Urubamba Valleys. Altitude of bench mark approx. 15748 ft.

6. — « Huaytapayana Glacier », in the « Huaytapayana Mountain Range » ($12^{\circ} 6'$ South $75^{\circ} 10'$ West) NE of the City of Huanayo, situated in the eastern Cordillera. Altitude of bench mark approx. 15420 ft.

7. — « Tuyujutu Glacier » ($12^{\circ} 5'$ South 76° West) on the south-eastern slopes of « Mount Tuyujutu » situated in the Western Cordillera. Altitude of bench mark approx. 16076 ft.

8. — « Yanasinga Glacier » ($11^{\circ} 35'$ South $76^{\circ} 10'$ West) located in the « Western Cordillera », east of Lima. Altitude of bench mark approx. 15802 ft. (see : Bol. del Inst. Geol. del Perú, No. 5, Lima 1946, Phot. page 16).

9. — « Huagoruncho Glacier » ($10^{\circ} 30'$ South $75^{\circ} 50'$ West) at the western slopes of « Mount Huagoruncho » located in the Eastern Cordillera in Central Peru, east of the town of Cerro de Pasco. Altitude of bench mark approx. 16732 ft.

Of all these glaciers mentioned above only three have been remeasured in the recent past :

« San Francisco Glacier » is one of the few which may be considered still a valley glacier. Its tongue descends more than 1000 m from the area of nourishment. A bench mark was set in front of the terminus in Sept. 1945 by the present writer and revisited by Arnold Heim in March 1947. According to these investigations the glacier terminus had receded horizontally an amount approx. 16 m in only 19 months. This is not astonishing if we remember that its marginal recession had been almost

9 m in only 5 month and 35.5 m in more or less 5 years. (see : V. Oppenheim y Hans J. Spann, « Investigaciones Glaciológicas en el Perú », Bol. 5 del Instituto Geológico des Perú, page 27). Although there are no data on volume shrinkage it is obvious that the glacier is waning away rapidly.

In Sept. 1945 a bench mark was established at the ice-front of « Incachiriasca Glacier », a transection type, and revisited eleven month later. A recession of 2 m was calculated which may be due to recession, but more likely to lack of exactness in our measuring methods.

The behavior of « Yanasinga Glacier », a small hanging icefield nearly reduced to its cirque and once a tributary of a huge glacier which occupied the Morococha Valley, has been summarized by several observers (Oppenheim 1944; Spann 1945; Spann and Petersen 1947) indicating a recession of about 14 m during that time.

It is to be mentioned that our methods of measuring (with tapemeasure and Brunton Compass only) are not at all satisfactory and the data obtained lack exactness: on the other hand surface ablation and volume shrinkage in general are not taken into consideration. It seems highly desirable that not only the ice-fronts (glacier terminus) but also the area of nourishment is mapped in a convenient scale. The planetable survey of the ice-front should be repeated every two years in order to accumulate enough data to record glacier oscillation in considerable detail.

Lima, April, 1 1948.

8) **DEGLACIATION IN COLUMBIA**

by

OPPENHEIM

At the kind request of Dr. J. A. Broggi, I have pleasure to report on my observations on the ice and snow conditions in the Cordillera Oriental of Columbia.

Although the writer has ascended the several snow peaks in Colombia, such as the Sierra Nevada de Snata Marta, Nevado del Ruiz and Nevado del Cocuy between the years 1939 and 1941, having everywhere observed an extensive withdrawal of the existing glaciers, it was only on the Nevado del Cocuy that actual comparison of the glaciers after a prolonged period of time could be made.

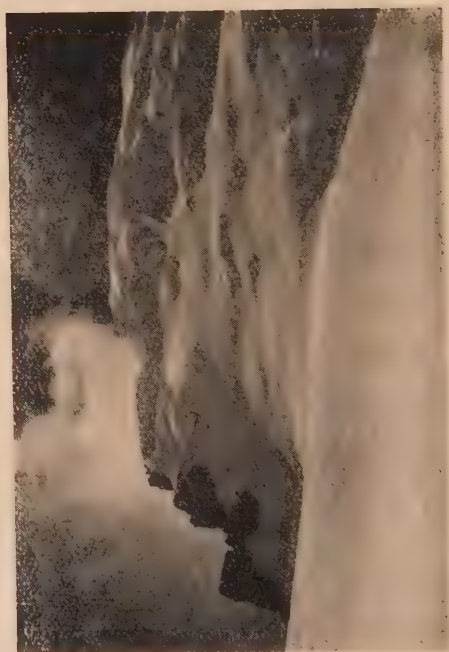
The Nevado del Cocuy was first ascended by the writer on December 25th 1939 having remained at the snow line until January 1st. Some observations on the conditions of the glaciers and numerous photographs were made at that occasion, while studying remains of pleistocene glaciations.

Another excursion to the Nevado del Cocuy at exactly the same site of Laguna de la Cueva was made on December 26th of 1947, remaining at the edge of the glaciers till December 29th. The position of the main glacier was observed at the edge of the Laguna de la Cueva at 4.600 meters.

The deglaciation of the Nevado del Cocuy since 1939 could be clearly observed by the writer. Although bench-marks were not put up at the time of the first excursion, comparative photographs show an unmistakable withdrawal of the front lobes of the glaciers in the last eight years. The withdrawal is however barely perceptible on the glaciers overhanging the steep mountain sides. Evidences of deglaciation are much more visible in the gentle slopes of the western flank of the Nevado. Here, while the glaciers in 1939 descended to the very edge of the Lake of Cueva. they were observed in December 1947 at a distance of some 20 to 30 meters from the lake shores and only in one spot was the main glacier descending directly into the lake. The vertical distance of withdrawal can this be estimated at about 7 to 10 meters, or an approximate rate of 1 meter per year could be reasonably suggested for the Nevado del Cocuy.

The attached photographs illustrate the conditions of the glacier lobes of the Nevado del Cocuy in December 1939 and December 1947.

Bogotá, April 1st 1948.





VI. — Rapports relatifs aux :
RECHERCHES SUR LES GLACIERS

REPORT ON THE PROGRESS

1) OF GLACIOLOGICAL RESEARCH IN GREAT BRITAIN

by

GERALD SELIGMAN

In England post-war preoccupations and particularly the task of academic reconstruction have prevented the resumption of glaciological research on any large scale. Nevertheless a start has been made.

The former *Association for Study of Snow and Ice* resumed activity in 1945 soon after the end of the war. Its name has been changed to the *British Glaciological Society*. The original *Association* was founded at the Congress of the I.U.G.G. at Edinburgh in 1936, largely through the inspiration given by the meetings of the then twin International Commissions of Snow and Glaciers. The revived Society retains its primary objects, « to encourage research on, and to stimulate interest in, the practical and scientific problems of snow and ice ». It is open to glaciologists and any others of all nationalities having genuine interest in any aspects of the subject.

The Society now publishes the *Journal of Glaciology* in order to reinforce these aims. Its purpose to be international has already had a measure of success; the circulation of over 475* is still increasing and extends to most parts of the world. Facilities are also provided for the *Journal* to be sent to institutions and to persons who are not members of the Society.

The Society meets about four times a year and papers on glaciological subjects are read and discussed. It has naturally taken a little time to produce papers on new research since practically all work on the subject had ceased during the war years. However investigations were made on the distribution of sea ice in the Arctic seas during this period by A.R. Glen and R. Moss, and the glaciers of central Africa were surveyed by Lieut. Commander P.C. Spink, R.N.

Of field work undertaken officially by the Society the most important so far is the resumption on a very comprehensive scale of the annual snowfall survey of Great Britain.

The Society has also organized a Glacier Physics Committee with headquarters at Cambridge. The Secretary is Dr. M. F. Perutz of the Cavendish Laboratory, Cambridge, well-known in glaciological circles for his crystallographic researches on ice at the Jungfrauoch in 1938. The Committee includes several well-known physicists, geophysicists and specialists in several subjects cognate to glaciology. It is at present elaborating a programme of laboratory and field research on glacier flow, and an expedition is now at the Jungfrauoch under the leadership of Dr. Perutz. Its preliminary aim is to test new and accelerated boring methods with a view to measuring speeds of flow deep down in the glacier and to correlate these with the shape of the bed obtained by seismic soundings. The personnel consists of members of the departments of Geophysics and of Geography at Cambridge and of the Geological Department of Reading University. The expedition is in close touch with the Swiss glaciological authorities through the friendly co-operation of Professor A. von Muralt of Bern and Professor R. Heafeli of the E. T. H. Zürich. The Forschungsinstitut Jungfrauoch is giving all possible assistance. Progress on this research, however, is not expected to be fast. I have just received a telegram from Dr. Perutz to say that they have had considerable success with a thermal boring tool, reaching a depth of 126 m. in a very short time. At this depth they reached either a very large boulder or the glacier bed. Inclinator tests showed that the bore hole was practically vertical.

Other bodies in England have also played a part in continuing glaciological research. In 1947 the Oxford University Exploration Club sent an expedition to Jan Mayen under A.J. Marshall and another to Iceland under F.H. Whitehead. The former made measurements on Sörbreen for comparison with the results obtained by the Imperial College of Science Expedition of 1938 under A. King. The latter worked on the Vatnajökull.

In 1946 and 1947 I.R. Menzies of Cambridge University led expeditions to Iceland and made extensive investigations of snow patches and snow patch erosion. Also in 1947 a large group from Cambridge consisting of members of the staff of the

* 630 April 1950.

University Geographical Department with research and senior students, under the leadership of W.V. Lewis, carried out research in Jotunheimen. Their chief object was to assess the nature of the flow of a glacier over its base and the consequent ability of glaciers to erode their beds. A small esker was found in process of formation and examined under such favourable conditions that much light promises to be thrown on the development of eskers.

During the same summer the writer visited many glaciers in the Swiss Alps making preliminary investigations for a large scale research on the growth of the ice crystal in glacier tongues. In 1948 he continued this work in the Norwegian mountains for the purposes of comparison with his Swiss results. Papers on these various researches are pending.

Several other expeditions with glaciological programmes are planned for this year. These include an expedition by the Durham University Exploration Society to Iceland to work in conjunction with Jón Eythorsson. Cambridge University has also sent a research party to Iceland and another to Norway, and a small party is also being sent from Leeds University to East Greenland.

A difficulty continually encountered by organisers of British expeditions to northern regions is the lack of transport. More expeditions would have started this year but for this difficulty. Help from Scandinavia by ships regularly plying to northern coasts has been gratefully accepted, but in one case an important expedition from St. Andrew's University to North West Greenland, planned for over a year, had to be abandoned owing to transport difficulties at the last moment. Another difficulty is the lack of trained and experienced glaciologists able to teach students and take charge of expeditions.

As is natural glaciological research in the Antarctic based on the home land has been slow to revive, but an important symposium on proposed Antarctic research was held in 1947 at Cambridge under the auspices of the *British Glaciological Society*. A considerable number of experts was present and a wide range of subjects awaiting investigation by forthcoming expeditions was discussed. This discussion has been reported in the *Journal of Glaciology* (Vol. 1, No. 3, 1948) and forms a valuable programme for future glaciological research, not only in the Antarctic, but in other parts of the world.

It will be seen from this report that the importance of glaciology is becoming recognized in a much wider field than formerly in Great Britain as well as in other countries. The research parties are still on a small scale and it may be some time before important results are obtained. Nevertheless it is clear that glaciological research in Great Britain is gathering momentum.

2) ON THE VARIATIONS OF WESTERN NORWEGIAN GLACIERS DURING THE LAST 200 YEARS

by

KNUT FÆGRI

The variations of glaciers are so obvious that they have been recognised wherever glaciers descend into areas inhabited or regularly visited by man. As a number of Western Norwegian glaciers end in the immediate neighbourhood of permanent habitation, the existence of glacier variations has been well known to the local population. Unfortunately, very little exact information is to be obtained from casual observations or from traditions, the latter being usually too vague as to time and locality. The data available from the time preceding the systematic yearly glacier measurements are therefore meagre, but in some cases they are sufficiently well defined to allow an analysis of the main features of that recession of our glaciers that started 200 years ago.

On the size of the glaciers in Western Norway before the 18th century

Very little is known about our glaciers from the time preceding the great advance of the 18th century. Unfortunately no investigations have been carried out on the conditions in Western Norway during the late-glacial period. In Eastern Scandinavia Mannerfelt (1941) has shown that the ice-sheets of the last glaciation passed through a dead-ice stage before disappearing, gradually melting away from the top downwards. Conditions in Western Norway differ in many respects from those further east: The horizontal distance from the high mountain nevéés down to sea-level is much shorter than in the east, the valleys are much steeper, and a glacier may therefore have « run out » as a living glacier even if it was dead in the sense of glacier economy, i.e. if it did not receive any nourishment from above. From casual observations we know that dead-ice formations do exist also in Western Norway (Mannerfelt l.c., Fægri 1939), we also know formations which seem to indicate that the ice in other places receded as a living glacier, but we do not know to what extent the ice behaved in one or the other way.

This is very unfortunate, as our principal means of identifying previous glacier stages will be by means of the moraines, provided these are datable. In front of the big Jostedalsbre outlet glaciers the moraines of the great advance that ended in 1750 (cp. below), are easily recognisable. There is in Jostedal a pronounced difference in the aspect of vegetation outside and inside these moraines (cp. Fægri 1933). The difference in soil development is even more pronounced. It must be admitted, though, that in the more genial climates of Fjærland and Olden on the S, resp. W side of the nevéé field these differences are less obvious than in Jostedal on the E side.

In front of the small Styggedalsbre in Hurrungene (cp. Ahlmann 1940s and previous publications) there is one distinct, obviously young moraine that constitutes a vegetational limit of the same type as the 1750 moraine in Jostedal. Outside this one, however, there is another moraine, which is equally distinct and also relatively young, but which does not constitute a similar limit, neither with regard to vegetation nor to soil development. Nothing similar is observed in Jostedal. The question of the existence or non-existence of moraines is hardly definitely settled (it is sometimes very difficult to distinguish between land-slides and moraine walls in those valleys), but no young moraines seem to be present outside the 1750 line. It thus appears that whereas the Jostedalsbre outlet glaciers in 1750 evidently reached further down than ever since the last glaciation (?), the Styggedalsbre moraines may register another young glacier transgression.

It is difficult to reconcile this conflicting evidence. One may presume that it has something to do with the size and morphology of the respective glaciers. The Styggedalsbre is a very small, self-contained unit, that reacts rapidly on climatic oscillations. The Jostedalsbre outlet glaciers come from a vast nevéé field, and they

are in themselves much longer than the Styggedalsbre. On account of this much greater volume we may presume that they will react more slowly. Thus it may be possible that the small glacier has reacted fully on a short, but pronounced change of climate which did, however, not last sufficiently long to influence the greater glaciers to the same extent. If that is the case, there must have been in a geologically speaking very near past another climatic deterioration similar to that which characterised the former half of the 18th century.

With this possible exception we possess no indications that our glaciers have recently been greater than they were about the middle of the 18th century, nor that the climatic conditions for glacier formation were more favorable than during that period.

Arrows with their wooden shafts and in some cases even their feather vanes have been found on the surface of vanishing snow-patches in the mountains of Central Norway (cp. references in Fægri 1938). As their state of preservation clearly shows that these arrows have not been exposed to the air since they were embedded in the « permanent » snow centuries ago, they furnish very valuable indications as to the variations in size of the « permanent » snow-patches and thus also of the yearly snow surplus in the area. The oldest types found date back to the migration age, thus indicating that the snow-patches of a few years ago were of the same size as those of the migration period and that during the intervening 1400-1500 years they were greater. In other words: the yearly snow surplus in the thirties was smaller than ever since the migration period, but at that time it must have been correspondingly small. There is some reason to believe that the preceding archaeological period, the younger Roman iron age, was characterised by a genial climate, and that not only the snow-patches, but also the glaciers were at that time as small as — or most probably much smaller than — they are to-day. We may even question if any living glaciers existed in Western Norway then.

The 18th century advance and the recession

Apart from the very vague indications mentioned above, we possess very little evidence about the state of our glaciers before ca. 1700. Some time during the late 17th century a very strong glacial advance made itself felt: the glaciers came down into the valleys, covered previous pasturages, and even destroyed farms and houses. Thanks to this last-mentioned effect we possess some evidence of the extent of the damages as land-taxation protocols reflect the diminishing value of the affected farms. This evidence has been collected long ago by Rekstad (1901, 1904).

It is, of course, very difficult to establish the extent of the glaciers before the advance. Probably they were of about the same order as to-day. They then ad-

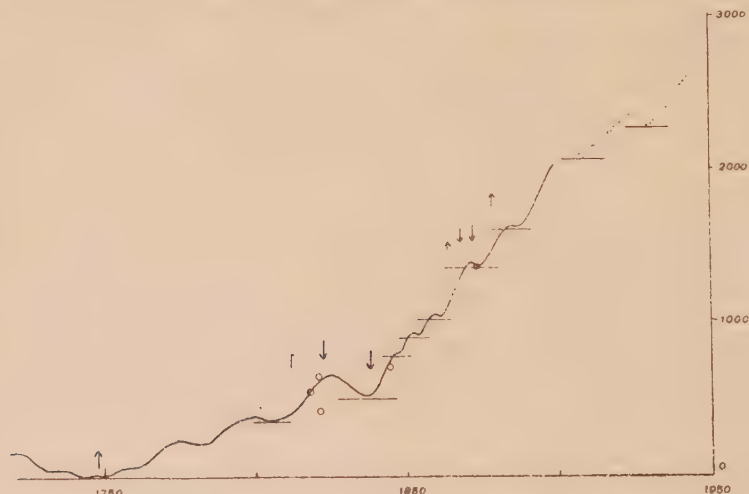


FIG. 1. Total recession of Nigardsbre since 1750. Short horizontal lines indicate moraines, rings: observations of the position of the glacier edge, and arrows observations of advance or recession. The full-drawn line is based upon these evidences, the dots indicate the measurements of the last 50 years. Notice the 1947 dot.

vanced some kilometers during the former half of the century. In 1750 the vicar of Jostedal, Foss, writes that the Nigardsbre had receded since 1748. Obviously the glaciers were oscillating for some time about the maximum stage before effective recession started, as is shown by the highly complex character of the 1750 moraines in front of Nigardsbre and other Jostedalsbre glaciers.

During the 19th century a number of scientists visited the more well-known glaciers and recorded their observations. By means of these records and by means of the intermediary moraines (which presumably must have been formed by secondary advances) I have for the Nigardsbre constructed the diagram shown in fig. 1, which gives the approximate history of the recession during the last 200 years. It will be seen that the recession till ca. 1850 was rather slight: even if the individual informations are perhaps not very reliable there can be no doubt about the position of the glacier edge in the 1820's. During the latter half of the 19th century and since 1930 the recession has been very rapid, whereas it was slower during the first third of this century.

The other glaciers are less well known, but there is every reason to suppose that their development has been on whole similar to that of Nigardsbre (cp. below).

The variations since 1900

In the years 1895 to 1908 Rekstad established regular yearly measurements of the variations of a considerable number of western Norwegian glaciers. Some of these series of yearly measurements have later been discontinued for some reason or other — during the last war contact was lost with some of the local assistants, some glaciers have receded so far as to be inaccessible, etc.

The measurements are carried out in the simplest way possible. A fixed point is marked on a great boulder or on the face of a rock. From this point the distance to the glacier edge is measured in a certain direction by means of a 100 metre string consisting of alternating 10 m pieces of brass and German silver. The distance from the last soldering point to the edge of the glacier is measured by means of an ordinary metre stick. These measurements are carried out by local assistants, most of which have « inherited » their jobs from some near relative who was instructed by Rekstad some 50 years ago. Considering that the measurements are thus carried out by unskilled personnel, and considering the rough terrain in front of the glaciers it is obvious that these measurements can make no claim for exactness. Besides the inevitable errors of measurement we must also take into consideration the possible variations in the direction of the measured distance from the fixed point to the glacier edge, and the possibility that the measured point is not representative for the

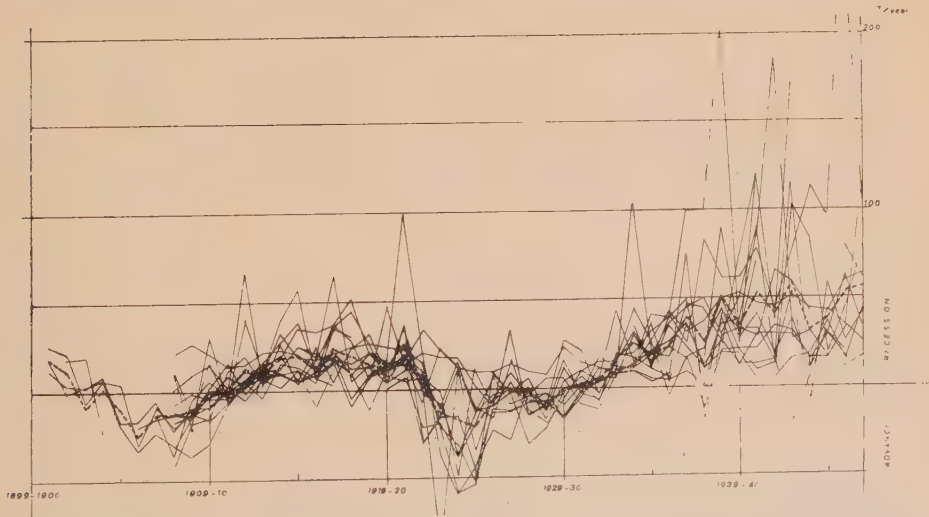


FIG. 2. Yearly movement of the edge of 15 Jostedalsbre outlet glaciers. The heavy dashes indicate the average.

glacier edge as a whole. The error thus is certainly of the order of some metres. However, results seem to indicate that on the whole these errors more or less compensate each other, and do not influence results too badly. It remains to be stated that because of the rapid recession during the last years, the distances from the fixed points to the glaciers are in many cases too long, which causes a corresponding increase of the errors. However, as long as the situation is as unstable as it is today, it is very difficult to do anything about this by changing fixed points, etc. It is very much to be hoped that these measurements can be taken up by some one who can put them on a better footing. Personally I have considered it of prime importance not to break the continuity of Rekstad's series, which he left in my care in 1932. The principal value of these series is caused by their length, some of the glaciers having been under observation for almost 50 years now.

On the synchronicity of glacier variations

The results of the yearly measurements have been published by Rekstad and later by myself till the season 1938-39. Later results are summarised in the table at the end of this paper. At the outset the numbers seem rather chaotic even if some tendencies can easily be discovered. Fig. 2 summarises all measurements from the Jostedalsbre outlet glaciers. It is obvious that in spite of all irregularities there is a very good correspondence between the variations of individual glaciers.

Fig. 3 indicates the effect of the length of the outlet glaciers. It is seen that the curves of the shorter ones are less regular with greater extremes in both directions. Those of the longer ones are more consistent, especially that of Tunsberg-

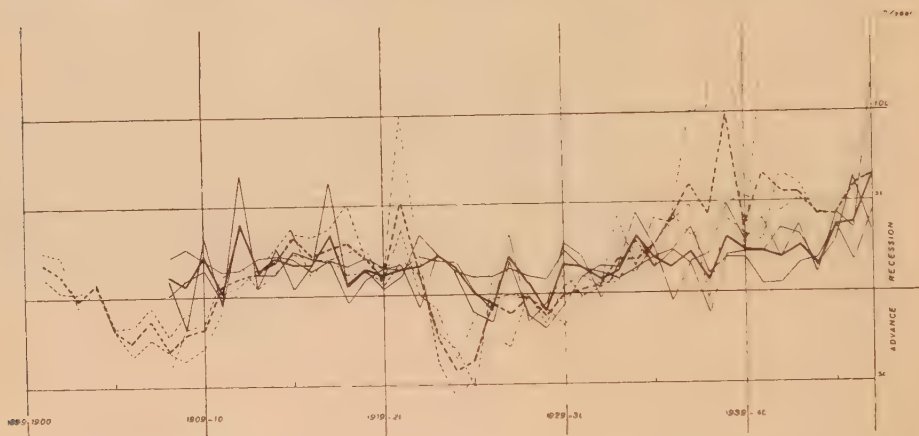


FIG. 3. Yearly movement of the edge of 3 short outlet glaciers from Jostedalsbre : Mjølkevollsbre, Briksdalsbre, and Bersetbre (dashes) and 3 long ones : Tunsbergdalsbre, Nigardsbre, and Lodalsbre (full-drawn). Heavy lines indicate the averages.

dalsbre, the longest one, which has advanced once only since measurements were started in 1907.

Of very great interest is the time lag of ca. two years between the advance of the short and the long glaciers at the outset of the advance period in the 1920's. Whereas the recession of all 3 short glaciers is strongly reduced during the season 1920-21, there is no consistent retardation of the recession among the long ones before 1922-23. This is, unfortunately, the only place in the diagram in which the time lag can be observed with any certainty. It should be noted, however, that whereas the main recession of the short glaciers takes place during 1922-25 and has definitely come to an end in 1925-26, that of the long ones takes place during 1925-26 and comes to an end in 1926-27, a lag of one year for the end of the first part of the advance period. It is also noteworthy that the secondary advance maximum in 1928-29 is in the case of the short glaciers rather small in relation to the first maximum, whereas it is even stronger than the first one in the longer glaciers.

There is no definite time lag in this part of the curves, but it seems as if there is some tendency to advance which has already been realised in the short glaciers whereas it is still latent in the long ones.

The glaciers dealt with so far are exposed towards S : Boyabre, Store and Vetle Suphellebre, towards SE : Austerdalsbre, Tunsbergdalsbre, towards E : Bersetbre, Nigardsbre, Fabergstolsbre, Lodalsbre, Stegaholtbre, and towards N : Mjølkevollsbre, Briksdalsbre, Abrekkebre, Kjendalsbre, and Bodalsbre. No systematic difference can be observed between the glaciers of different exposure.

Whereas there is an obvious correlation between the variations of the different outlet glaciers from the Jostedalsbre névée field notwithstanding their exposition, the correlation with variations of other glaciers is less apparent, cp. fig. 4. Folge-

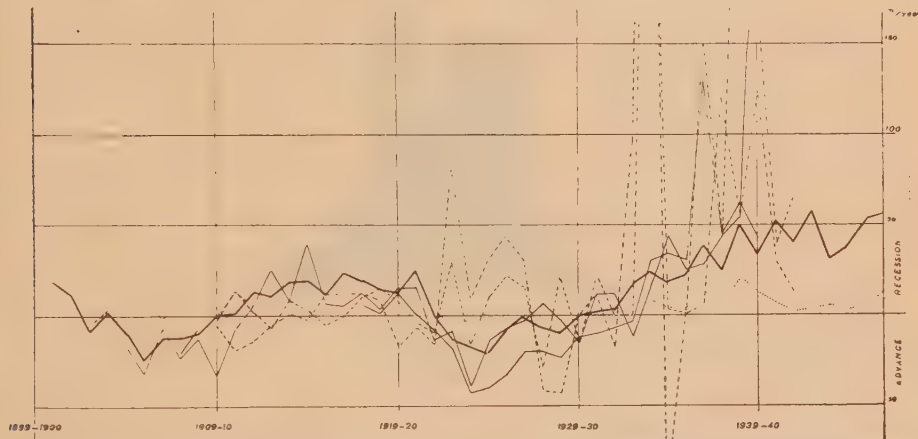


FIG. 4. Yearly movement of the edge of outlet glaciers from Folgefonno (full-drawn) and Svartisen (dashed) compared with the average for the Jostedalsbre outlet glaciers (heavy line) and Styggdalsbre (dots).

fonno is a névée field of the same type as Jostedalsbre, and is situated not far from the latter. The movements of its outlet glaciers are practically identical with those of the Jostedalsbre glaciers.

Svartisen also belongs to the same type, but is situated $4\frac{1}{2}^\circ$ further N. The curves for the 1930's exhibit strong disturbances and are difficult to compare with those of other glaciers. On the whole the Svartisen outlet glaciers show an accelerated recession during this period like those of the southern névée fields, but the details of the curves are not parallel. The pronounced advance period of the 1920's seems to begin somewhat earlier in the north, but it is interrupted by a very strong recession during the middle of the decade. It seems rather probable that the main features of the variation curves of the Svartisen and the southern glaciers will prove to be similar. However, our measurement series are not yet sufficiently long to prove this.

The small Styggdalsbre belongs to another type of glaciers, viz. the alpine. Its whole length, névée basin and all, is but 3 km, less than most of the outlets of Jostedalsbre. It is located close to Jostedalsbre, and it would have been of great interest to compare it with the outlets of the latter. Unfortunately the series is not continuous (cp. Ahlmann l.c.) and hardly permits any conclusions, even if there is an obvious correspondence with the Jostedalsbre curve.

The yearly period in glacier variations

Rekstad's original measurements were carried out once a year, usually in August-September. In 1935 monthly measurements were as an experiment started at Nigardsbre. Fig. 5 gives the average for the 12 years during which these observations have been carried out so far. It suffers from the great drawback that it is based upon a period of very strong recession, and cannot be taken as a typical representation of the variations of a western Norwegian glacier. It is noteworthy

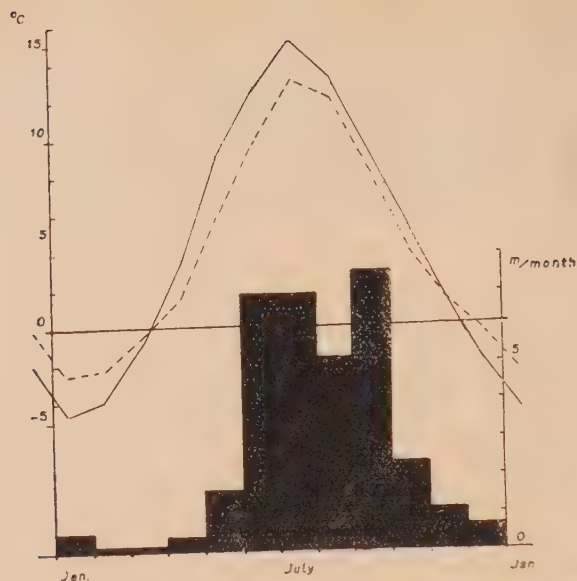


FIG. 5. Monthly recession of the Nigardsbreen October 1935 - July 1947 compared with mean monthly temperatures at Fjærland (full-drawn) and mean monthly temperature $+ 10^{\circ}$ at Fanaraken (dashes).

that there is an excess of recession even during the winter months. It must be taken into consideration, however, that the data are subject to significant errors of observation.

It is seen that the main recession takes place during the months June-September and the minimum during February-March. The best meteorological stations for comparison with the yearly variations of temperature are Fjærland at the immediate southern end of the Jostedalsglacier and the meteorological high mountain observatory at Fanaraken, 2062 m s.m. a little east of Jostedalsglacier. For the former I have calculated the mean monthly temperature for the period 1935-1944 by means of the data published in *Jahrbuch des norwegischen meteorologischen Instituts*. For the latter Spinnangr has (1947) published mean monthly temperatures for the period 1932-46. The data are strongly influenced by the abnormally cold winters of the 1940's. Because of frequent temperature inversions during winter the influence of the abnormally cold winter is obviously relatively stronger in a valley station like Fjærland than on the isolated mountain top of Fanaraken. I find it most probable that « normally » February is at Fjærland least as cold as January. This is actually the case if we calculate the mean for all observations years of this station : January -2.8 , February -3.2 . However, the Bergen temperature observations for the same years (1923-44) show a positive anomaly as compared with the long term normal of January : 0.7 , February : 0.5 . A similar anomaly in Fjærland would to a certain extent remove the difference between the two months in the 1923-44 observations.

However that may be, there is a definite time lag between the temperature curve and the monthly recession both with regard to maximum and — even more remarkable — minimum.

Causes of glacier variations

This theme is much too complicated to be dealt with in a short memoir like this, and has, besides, been discussed by a series of investigators already (cp. Ahlmann 1947). It is, however, tempting to correlate the curves of glacier variations with temperature curves from western Norway, as Ahlmann has shown that with regard to glacier regime « temperature has been the mainly decisive factor » (1940 p. 125). We possess reliable temperature observations from Bergen (cp. Birkeland

1928, my thanks are due to Mr. Birkeland of the Norwegian meteorological office for supplementing these data up till present), midway between Folgefønno and Jostedalsbre. The parallelism of variations of the outlet glaciers from the two nevée fields warrants the use of this comparatively far-away station. Fig. 6 shows the glacier variations, viz. the above mentioned averages for the long and short outlet glaciers from Jostedalsbre compared with temperature at Bergen. The middle temperatures of the three seasons spring (March-May), summer (June-August), and autumn (September-November) are given separately. In order to reduce the short-period disturbances the curves have been smoothed by the calculation of consecutive 3 years' averages. As a cumulative temperature effect is highly probable, such a procedure is certainly justified. It should be noticed that the averages are set down under the *last* year (i.e. the 1901-03 average is considered as belonging to 1903, not 1902 as usual), as the glaciers cannot, of course, react on

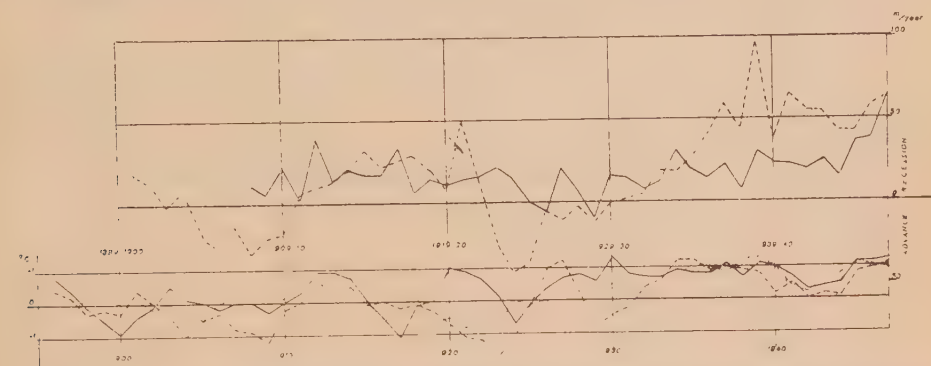


FIG. 6. Recession of short (dashes) and long (full-drawn) outlet glaciers from Jostedalsbre (upper graph) and 3 years average seasonal temperature anomaly at Bergen (lower graph, spring : full-drawn, summer : dashes, autumn : dots).

future temperatures. The graph does not aim at giving more than a first approximation. The calculation of temperature curves has been carried out in the simplest way possible without any of the many reduction factors which it may seem desirable to investigate further (e.g. an average of the type $m_a = \frac{c + 2b + 3a}{6}$).

Our glacier measurement series are not yet sufficiently long to promise very much from a mathematical treatment of the two graphs of fig. 6, and I don't think that it will at present reward the labour involved in calculation of coefficients of regression.

It is, however, obvious that the advance period of the 1920's is correlated with the period of negative temperature anomalies that started in 1916. It is seen that the rate of recession slowed down almost instantly, even if the real advance did not come until 1922 (short glaciers) or 1924 (long ones). The deviating spring temperature curves of the yearly 1920's complicate the picture and makes it difficult to evaluate the length of a contingent time lag. It is even less profitable to discuss the imperfectly known advance period of 1905-11, even if there is an obvious correlation with negative temperature anomalies. The recession period of the 1910's has no distinct parallel in the temperature records even if there is a positive anomaly in the spring and summer temperatures for a few years. The high positive anomalies of the 1930's are beautifully correlated with the glacier variations and no systematic time lag can be observed between the 3 years' average and glacier variations. The lower positive anomalies of the early 1940's do not seem to have influenced the rate of recession very much, even if there is a distinct acceleration during the last years when positive anomalies were again high.

Bergen, May, 1948.

REFERENCES

- AHLMANN, H. Wson. 1940 : The Styggedal glacier in Jotunheim, Norway. — *Geogr. annaler.* 1940, 95.
 1947 : Den nutida klimatfluktuationen och dess utforskande. — *Norsk geogr. tidsskr.* 11, 290.
- FOSS, M. 1750 : Justedalens kortelige beskrivelse. Edited posthumusly by J. Berg. *Magazin f. Danmarks og Norges topographiske, oekonomiske og statistiske Beskrivelse.* 2 (1802-03), 1.
- FÆGRI, K. 1933 : Ueber die Längenvariationen einiger Gletscher des Jostedalsbre und die dadurch bedingten Pflanzensukzessionen. — *Bergens mus. arbok* 1933. *Naturv. rk. nr.* 7.
 1938 : Forandringer ved norske breer 1936-37. *Ibid.* 1938 nr. 4.
 1939 : Features of the late-glacial development of northern Jæren and the adjacent highlands. — *Norsk geogr. tidsskr.* 7, 121.
- MANNERFELT, C. 1945 : Nagra glacialmorfologiska formelement och deras vittnesbörd om inlandsisens avsmålningsmekanik i svensk och norsk fjällterräng. — *Geogr. annaler* 1945, 1.
- REKSTAD, J. 1901 : Iagttagelser fra bræer i Sogn og Nordfjord. — *Norges geol. unders., aarbog* 1902, nr. 3.
 1904 : Fra Jostedalsbræen. — *Bergens mus. aarbog* 1904, nr. 1.
- SPINNANGR, F. 1947 : Fanaraken meteorologiske observatorium. — *Naturen* 71, 97.

Table. Yearly variations of some western and northern Norwegian glaciers.

Jostedalsbre:						
Years	Boyabre	Store Sup- hellebre	Vetle Sup- hellebre	Auster- dalsbre	Tunsberg- dalsbre	Bersetbre
1900-01	(— 80½)	(+ 6	(— 37)	—		
-02	1899-03)	— 20	1899-02)	—	(34½	(—112½
-03		— 20	+ 66	—	1900-03)	1899-03)
-04	— 13½	+ 24		—		
-05				—	(± 0	(+ 61
	(+ 83½	(+ 13½	(+ 52		1903-07)	1903-07)
1905-06	1904-06)	1904-06)	1903-08)	+ 20		
-07	+ 13	± 0?		+ 15		
-08	+ 36	+ 17		+ 13½	— 1	+ 42
-09	+ 5	+ 13	+ 21	+ 14	— 9½	+ 17½
-10	— 6	+ 8	(+ 9)	— 16	— 15	+ 15
1910-11	+ 2	— 3		— 1½	+ 3	— 11
-12	— 4	+ 5	— 16	— 3	— 67	— 20
-13	— 12	— 7	— 9	— 8	— 3	— 4½
-14	— 40	— 2	— 32	— 9½	— 26	— 13
-15	— 57	— 11	— 20	— 11½	— 4	— 24½
1915-16	— 7	+ 9	— 7	— 4½	— 17	— 21
-17	— 3	— 12	— 13	— 35	— 64	— 20½
-18				— 30½	— 5	— 24½
-19	(— 10)	(— 11)	(— 6)	—	— 14	— 26½
-20	— 11	— 1	— 17	—	— 4	— 5½
1920-21	— 37	— 7	— 33	—	— 10½	— 19½
-22	+ 30	+ 21	+ 39	—	— 33½	— 11½
-23	+ 15	+ 19	+ 11	—	— 23	+ 21
-24	+ 15	+ 38	+ 36	—	— 13	+ 30½
-25	+ 45	+ 19	+ 21	—	— 10	+ 52½
1925-26	+ 13	— 8	+ 15	—	— 10	— 8
-27	+ 13	+ 1½	+ 2	—	— 14½	+ 2
-28	+ 8	+ 3	— 6	—	— 10	+ 2½
-29	+ 3	+ 11	+ 13	—	— 8	+ 8
-30	+ 17	— 16	+ 18	—	— 23	+ 17
1930-31	+ 5	+ 3	+ 5	—	— 14½	+ 1½
-32	— 11	— 17	— 7	—	— 12	— 6
-33	— 10	— 12	— 18	—	— 8	— 5½
-34	— 10	— 11	— 14	— 35½	— 13½	— 28½
-35	— 18	— 37	— 15	— 11½	— 20	— 13½
1935-36	— 35	— 3	— 43	— 36	— 26	— 39½
-37	— 45	— 35	— 40	— 23	— 36½	— 49
-38	— 50	— 13	— 15	— 11	— 10½	+ 4
-39	— 30	— 27	— 90	— 36	— 19½	— 41
-40	— 50	— 13	— 30	— 38	— 19	— 32½
1940-41	— 90	— 13	— 70	— 86	— 4½	— 29
-42	— 15	— 15	—	— 187	— 5	— 66
-43	— 115	+ 20	—	— 27	— 16	— 60
-44	— 10	— 5	—	— 31	— 18½	— 42
-45	— 30	— 32	—	— 16	— 36½	—
1945-46	— 20	— 17	—	— 11	— 17	—
-47	—	—	—	— 17	— 43½	—

Numbers in parentheses comprise more than one year. They are not used for the graphs. Interval indicated where necessary. Lires : no measurement.

Nigards- bre	Foberg- stólsbre	Stegaholt- bre	Lodalsbre	Aabrekke- bre	Briksdals- bre	Mjólke- vollsbre
		—		— 18	— 12½	— 25½
(— 73½ 1899-03)	(— 22 1899-03)	—	(— 74½ 1899-03)	+ 1½	— 3	— 23
		—		— 3	— 2½	+ 6
		—		— 7	— 9	— 6
(+ 9 1903-07)	(— 35 1903-07)	(+ 9 1903-07)	(— 72 1903-07)	— 5	+ 17	+ 19
				+ 30	+ 16½	+ 33
				+ 8	+ 5	+ 23
— 10	— 10½	+ 11	— 21½	+ 20	+ 22	+ 30
+ 18½	+ 11	+ 14	— 27	+ 13	+ 9	+ 36
— 31	+ 3½	+ 2½	— 21	+ 15½	+ 9	+ 29
+ 5	— 8	— 1½	— 14	+ 7	— ½	+ 7
— 40	— 4	— 11½	— 15	— 5	— 6	— 4
— 11½	— 13½	— 9½	— 22	+ 3½	— 16	— 15
— 12½	— 9½	— 10	— 23	— 15½	— 27½	— 18½
— 25	— 5	— 5	— 21	— 13	— 38½	— 35
— 20	— 10½	— 9	— 14	— 7	— 11	— 33
— 19	— 17	— 14½	— 19½	— 37½	— 22	— 37
— 15½	— 9½	+ 9	+ 4½	— 28	— 12½	— 51½
— 21	+ 11	— 7½	— 9	— 6	— 13	— 21½
— 14½	— 10	— 11	— 15	— 47	— 8	— 15
— 16½	— 16	— 13½	— 14½	— 16	— 36	— 100
+ 7	— 4½	— 5½	— 21	— 8	± 0	— 21
— 23	+ 81	+ 15½	— 18½	+ 1	+ 8	+ 37½
— 13	— 13	— 12	— 18	+ 48	+ 36	+ 58
+ 10	+ 13½	+ 34	— 1½	— 10½	+ 13	+ 44
+ 16	— 6	+ 5½	+ 9	+ 11	+ 3	+ 23
— 16	+ 8½	— 9	— 33	— 4	+ 1	+ 29
+ 12	+ 3½	+ 2½	+ 15½	+ 14	+ 2	— 1
+ 20	+ 2½	— 6½	+ 7	+ 9	+ 12	+ 17½
+ 6	+ 9½	— 6½	— 27½	+ 2½	— 11½	— 4
— 9	— 1½	+ 9½	— 21	— 7	+ 2	— 10
— 15	— 2½	+ 12	+ 3½	+ 3	— 3	— 4
— 15	— 8½	— 8	— 14½	— 17½	— 9½	— 44
— 45	— 9	— 10	— 32	— 19½	— 12	— 14½
— 25	— 3	— 5	— 14	— 32	— 27	— 41
+ 5	— 7½	— 4	— 23	— 39½	— 43	— 39
— 17	— 75	— 9	— 16	— 25	— 31	— 100
— 21½	+ 18½	— 1	+ 12	— 83½	— 37	— 100½
— 49½	— 29	— 39	— 20½	— 61½	— 50	— 202½
— 28	— 33	— 12	— 22	— 61½	— 53	— 23
— 41	— 32½	— 30	— 22	— 78	— 47	— 120
— 19	— 17½	— 30	— 35	— 44	— 45	—
— 38	— 53	— 29	— 32	— 79½	— 51½	—
— 10	+ 7½	— 16	— 14½	— 113½	— 44	—
— 43	— 59	— 16	— 28½	— 96½	— 42½	—
— 35	— 26	— 34	— 63½	— 254	— 60	—
— 113	— 42	— 25	— 33½	— 105	— 64	—

Folgefønno :			Svartisen :			Stygge-
Kjendalsbre	Bodalsbre	Buarbre	Bondhusbre	Engabre	Fonndalsbre	dalsbre
(— 6 1900-07)	(+ 30 1900-07)	(—150 1900-04 — ½	— — (+ 5) + 16½	— — —	— — —	— — + 5½ — 3 — 4
				(+ ca 100 1903-09)	(+ ca 60 1903-09)	— 12½ + 2½ — 8½
+ 31 + 13 — 2	+ 7½ + 1 + 4	+ 21 + 7 (+ 50)	+ 23½ + 13 + 34	+ 3	+ 5	— 9 ± 0
— 16 + 1½ — 10½ — 32½ — 18½	+ 4 — 1½ — 12½ — 20 — 6	— 18 (— 57)	+ 7 — 5 — 25½ — 8 — 40	— 13 — 2 + 7 — 9 — 3	— 19½ + 13½ + 5 — 2 + 2	— 14 — 5 — — —
— 17½ — 36 — 45 — 26½ — 24½	— 24½ — 5 — 23½ — 31½ — 8½	— — — 6 — 1 — 12	— 6 — 5½ — 12½ — 4 — 15	+ 5 ± 0 — 10 — 15 — 8	— 12 — 13 — 13 — 9 + 17	— — — — —
— 22½ — 15½ + 20½ + 57 + 53½	— 5 — 6½ + 10½ — 15 + 12	— 1 + 8 + 18 + 43 + 40	— 16 + 13½ + 8½ + 39 + 14	+ 4 + 10½ — 30 + 16 — 10	+ 6 + 16 — 31 — 9 — 31	± 0 ± 0 + 2 + 3 — 1½
+ 11 + 12 + 7½ + 6½ + 16	+ 9½ + 2 + 31 + 22 + 1	+ 33 + 20 + 20 + 23 + 12	+ 7 + 3 — 6½ + 3 + 15½	— 22 — 16 + 29 — 22 + 15	— 44 — 30 + 42 + 43 ± 0	(— 35½)
+ 3 + 6½ — 11½ — 105 — 19½	— 1½ + 7½ — 19½ — 17 — 10½	+ 10 + 7 + 4 — 30 — 34	— 11 — 12 + 11 — 14½ — 44	— 23 ± 0 ± 0 — 332 + 104	— 12 + 18 — 96 — 561 — 3	— 9½ — 4
— 17½ — 46½ — 44½ — 24 — 74	— 6 — — 25 — 44 — 20	— 31 — 129 — 45 — 62 — 43	— 25½ — 28½ — 44 — 54 — ca280	— 4 — 150 — 43½ — 481 — 172	— 1 — 6 — 120 — 55 — 124	— 4 — 12½ — 5 — 20½ — 13
— 120½ — 37½ — 174 (— 331½)	— 10 — 13 — 104 — 83½ — 19	— — — — —	— — — — —	— 40 — 65 — — —	— 29 — 14 — — —	— 7 — 2 — 3½ — 5½ — 3
— 80 — 62	— 35 — 70½	— —	— —	— —	— —	— 5 — 13

3) GLACIER VARIATIONS AND CLIMATIC FLUCTUATIONS IN BRITAIN

by

GORDON MAULEY

Largely as a result of Scandinavian work it has become possible in recent years to correlate the behaviour of glaciers with the trend of monthly and annual means of temperatures at Stockholm during the past 200 years. (Ahlmann, Angstrom) While some have questioned the validity of 18th century records it is particularly interesting to note that other long series of monthly means eliminating the possible effects of urban exposure have now become available for N.W. England and Holland (Fegedal). The trend of the independent English curves during the past 200 years is so much in accordance with that of Sweden as to give further confidence not only in the reality but also the amount and time of development of the present secular change (Maulen, Labriijn).

It is hoped to extend the English records at least back to 1715 and thus to reinforce Labriijn's figures (1706 onward) and I am not without hope of devising further methods of estimating the extent and character of the temperature fluctuations for the later 17th century. If this can be done we shall have view of the temperature trend in N. W. Europe covering Thorarinsson's curve of glacier oscillation in Iceland, and about half of Matthes' « Little Ice Age ». From the nature of this curve it may be possible to estimate the character of the events initiating the « Little Ice Age », and why it appears to have begun in N.W. Europe about 200 years before the great advance of the Alpine glaciers. The series of events about 1300-1350 in England is at present attracting attention; but evidence must be gathered very slowly from the most diverse sources. It may be added that such evidence as we possess with regard to the persistence of the semi-permanent snow-beds in Scotland agrees very well the behaviour of glaciers in Iceland and Norway.

In discussing glacier behaviour and climatic fluctuations it is, however, necessary to take a very general view, and to disregard individual glaciers, or the results of individual seasons: further we are often using very imperfect observations. One must think in terms of whole countries and of groups of years, and one must try to bear in mind climatic agencies which will take effect over groups of years, and even whole centuries. For N.W. Europe one of the fundamental agencies of this kind must surely be sought in the extent off Greenland of the melt-water from the Arctic ice.

Ahlmann has shown that a slight rise or fall in spring and autumn temperatures is extremely effective with regard to the behaviour of glaciers in Norway. The English instrumental record for the period immediately preceding one of the dates of maximum post-glacial advance (about 1745) suggest that the occurrence of a small group of six or seven cold years, in which cold spring and autumn years were especially notable, is sufficient to lead to sudden and marked glacier advance. (Werneskiold, Faegri, Ahlmann). While the fluctuations of spring temperatures in Britain are not as large as in midwinter or midsummer, the rises and falls compare remarkably well with Thorarinsson's curve of glacier behaviour in Iceland. Bearing in mind that in an insular climate in May and October the temperature of the sea-surface is more important than that of the land it can reasonably be suggested that one of the most important factors with regard to the temperature variations in England, and with regard to glacier behaviour in Norway and Iceland is the location of the border of the melt-water from the Polar ice. Careful analysis of the small variation in mean temperature at stations ranging from Stykkisholm and Lofoten to Plymouth, in the regions adjacent to an open sea where maritime polar air masses are dominant, might make it possible to estimate the extent of the movement of this « Arctic convergence ». (Eriksson)

In the late 18th century the mean temperature for *warm* winter months in England, Holland or Sweden was very much the same as now. Remembering the properties of air masses it would therefore appear that the sea temperatures off the coasts of S.W. Europe must have been much the same as at present. Hence for the climate as a whole to allow of glacial advance it is hard to resist the conclusion that the maritime-polar air must have had a shorter travel over the warm water. Apart from the possibility of increased wind speed the conclusion seems to be that there was a greater spread of Arctic 'water.

With regard to earlier post-glacial time, recent work in England has confirmed the occurrence of the Allerød oscillation already established in Denmark and Norway. The extent and character of this oscillation affecting N.W. Europe might well be explained by adopting Angstrom's theory of a period of weak atmospheric circulation at a time when the sea off N.W. Europe was still rather cold, giving near the coast a mean July temperature of the order of 10° C in N. Ireland, and S.W. Norway (Pennington, Fægri, Godwin). For a later period in the climatic optimum it has been shown that in Denmark the summer temperatures were little warmer than the present values although the winters were milder, a conclusion which might well be explained by assuming the vigour of the atmospheric circulation to be much the same as now, but with considerably less « Arctic water » in the Norwegian Sea. (Iversen). Variations in the intensity of the circulation will explain the features of the Sub-Boreal, and (following Brooks) a rapid re-advance of the Arctic water at a time of vigorous circulation will explain the climatic deterioration of the Early Iron Age while retaining fairly mild winters near the coasts. They may also explain the recurrence-surfaces to which Granlund and Faegri have called attention.

These suggestions are made to illustrate my view that a careful study of all the evidence given by our instrumental records covering the past 300 years might go far towards elucidating the causes of the Pleistocene Ice Age in N.W. Europe. In particular we should take into account the views of Wager that the Greenland Ice Cap was already in existence in the Miocene (Wager, 1933). It may be very bold : but I should like to make the suggestion here to those better qualified than I am, that we should study the rate at which the cooling of the seas would then proceed in order to explain the lapse of time before the peripheral ice caps of Scandinavia and Laurentia were established. It would be interesting to have further knowledge of the time of uplift of the Scandinavian mountain system to a level at which glaciation would be initiated. For, as Dahl has shown, it is necessary to have a reasonably open sea and relatively mild temperatures on the N.W. coast of Norway at the height of the last phase of glaciation (Dahl). But it is similarly necessary to have an open sea close to Ireland, yet with a surface temperature which could scarcely be more than 5° C, in as much as the firn-line in N. England in midsummer can scarcely have exceeded 600 m above sea level, disregarding for the moment the effects of isostasy which ought not to be large in Britain (Flint). Moreover, the Atlantic off Scotland and N.W. Ireland must have contained a great deal of floating ice. This leads us to the conclusion that whereas pack-ice did not permanently cover the Norwegian sea the spread of cool surface water may even have extended to the parallel of 50° N. from its present position N. of Iceland, a view supported by Sir G. Simpson (Simpson).

Yet, at the height of glaciation Greenland, bordered by extensive pack-ice, must have been starved of precipitation, as the Antarctic is at present. Presumably, therefore, isostatic recovery of Greenland would in due time set in. If we accept Umbgrove's views of the relationship of large scale glaciation to the periods of orogenic uplift it remains for geophysicists to explore the interesting possibility of explaining the multiple glaciation of the peripheral land masses in terms of successive isostatic depressions and recoveries, following a long period during which the seas were cooled as a result of the mid-Tertiary elevation of Greenland. In such an event, based on Wager's views and on such indications as we have from the Antarctic, we should be able to equate in broad terms the length of the glaciation following the Alpine or orogeny with that suggested for the Upper Palaeozoic glaciation, namely 25 million years.

No doubt this appears a tremendous leap from the contemplation of the instrumental observations of the 18th century kept by our ancestors in various parts of N.W. Europe. But I believe it is for a geographer to open up material for discussion by those better qualified to make more detailed studies. Minor studies must be made of the reasons for the small groups of colder years within the record. I hope these remarks will have served to direct attention to the value of such studies as a key to interpretation, and the importance of linking and comparing such series, now that it has been shown by the Scandinavian glaciologists that we have so recently emerged from a period of maximum post-glacial advance. Moreover, it will be a fitting tribute to the great tradition of the observational scientists of N. W. Europe, whether in the realms of geology, botany, or meteorology; from John Olafsson of Sweden and Sveinn Pálsson of Iceland; Wargentin of Uppsala, Blytt and Sernander, Warming and Ostenfeld, Huyghens and Musschenbroek, Hjort and Sars. We all live round the North Sea and we all know maritime-polar air is, what, ever since our ancestors sailed in it, built their windmills and grew their corn.

4) OBSERVATIONS IN THE FIRN AND ABLATION REGIONS OF THE GREAT ALETSCHE GLACIER

by

R. HAEFELI and P. KASSER,

Laboratory for Hydraulic Research and Soil Mechanics of the Swiss Federal Institute of Technology, Zurich.
Published by the Glacier Commission of the Swiss Society for the Advancement of Science

Contents :

- I. *Introduction*
- II. *Climate, Firn Augmentation and Firn Temperatures*
 - 1) Meteorological Data
 - 2) Firn Augmentation and its Water Equivalent
 - 3) Firn Temperatures
- III. *Height Variations of the Glacier Surface*
 - 1) Continual Height Variations of Point 3 (1941-1947)
 - 2) Annual Height Variations of Firn Surface (1851-1947)
 - 3) Consequences of Glacier Shrinkage
- IV. *Flow Velocities at the Surface and in the Interior of Glacier*
 - 1) Creep Diagram above Bergschrund
 - 2) Conditions of Motion below Bergschrund
 - 3) Various Observations in the Ablation Region

List of Figures :

- 1) Map of Glacier Site
- 2) Photograph of Firn Region
- 3) Precipitation, Firn Augmentation and Air Temperatures
- 4) Heights of Snow during the Hydrological Year 1944/45 (Schneehöhen im hydrologischen Jahr 1944/45)
- 5) Driving Diagrams and Firn Temperatures
- 6) Temperature Conditions at a Firn Basin and its Subsoil (diagrammatic)
- 7) Continual Height Variations of the Firn Surface at Point 3
- 8) Annual Height Variations of Firn Surface near Point 3 (Profiles 1 and 2)
- 9) Height Variations of Firn Surface at Profile Firn
- 10) Creep Diagram above Bergschrund
- 11) Effective Shift of a Point of the Firn Surface
- 12) Distribution of Surface Velocities (Situation)
- 13) Formation of Tension Cracks (Schema)
- 14) Photograph of Concordiaplatz (from Kamm)

I. Introduction

In order to obtain as complete a view as possible of the glaciological appearance and their functional cooperation in the organism of the glacier, these studies aim at a coordination of methods of investigation of a single, sufficiently large glacier. These methods are today made available by certain branches. Contrary to many earlier investigations, which neglected the origin of a glacier due to its inaccessibility the firn region, where all the phenomena of glaciology originate, should be the centre of our present investigations. The Great Aletsch Glacier has been chosen as object of investigation, in the firn Region of which the Jungfrau Scientific Station — near the terminus of the Jungfrau Railway — represents an ideal base for glaciological research.

After the middle-age advance of glaciers in 1600, the Alpine glaciers reached a new noteworthy excessive advance again in the middle of last century. It is of

special benefit to glaciology that the first topographical survey of the Great Glacier Aletsch originates precisely from this era, containing levels of different points of the glacier surface. At that time the Stingelin map represented a fine piece of work (1851). For the first time in 1869 the speed of the glacier surface was measured by Grad and Dupré. The Federal Service of water power plants started water measurements at Massboden in 1921. From 1908 to 1914 Märjelsee was the subject of minute glaciological investigations by O. Lütischg [1] (1). Mothes and Sorge measured the depth of the ice at Concordiaplatz (max. about 790 m) in 1929 by means of echo-soundings [2] and in 1926 the Aletsch Glacier was surveyed photogrammetrically by the Federal Topographic Service. After the inauguration of

Fig.1 Basin of the Great Aletschglatscher



the Meteorological Station on Sphinx in 1937, Seligman and his assistants used the Jungfrauoch Scientific Station in 1938 for the first time as base for glaciological investigations [3]. Consequently starting with the Stingelin map, the accurate observation of the Great Aletsch Glacier leads back almost 100 years. The studies described in the following were developed originally by the Swiss Commission of Snow and Avalanche Research. Before deciding on a new station at Weisfluhjoch

(1) See Bibliography at the conclusion of the article.

there was the question of a possibly better site for the Snow and Avalanche Research Station in the Jungfrau region.

Based on simultaneous observations in both regions the answer was negative, and it was planned to carry out observations of the Jungfraufirn systematically during the whole following decade (1940-1950). After investigating the development of snowlayer and avalanches especially during the first 2 years of the above-mentioned period, the programme gradually concentrated on the following points, concerning the study of formation, motion and alteration of the glacier :

1) Continual observation and measurement of thickness of snowlayers, augmentation of firn and its water equivalent.

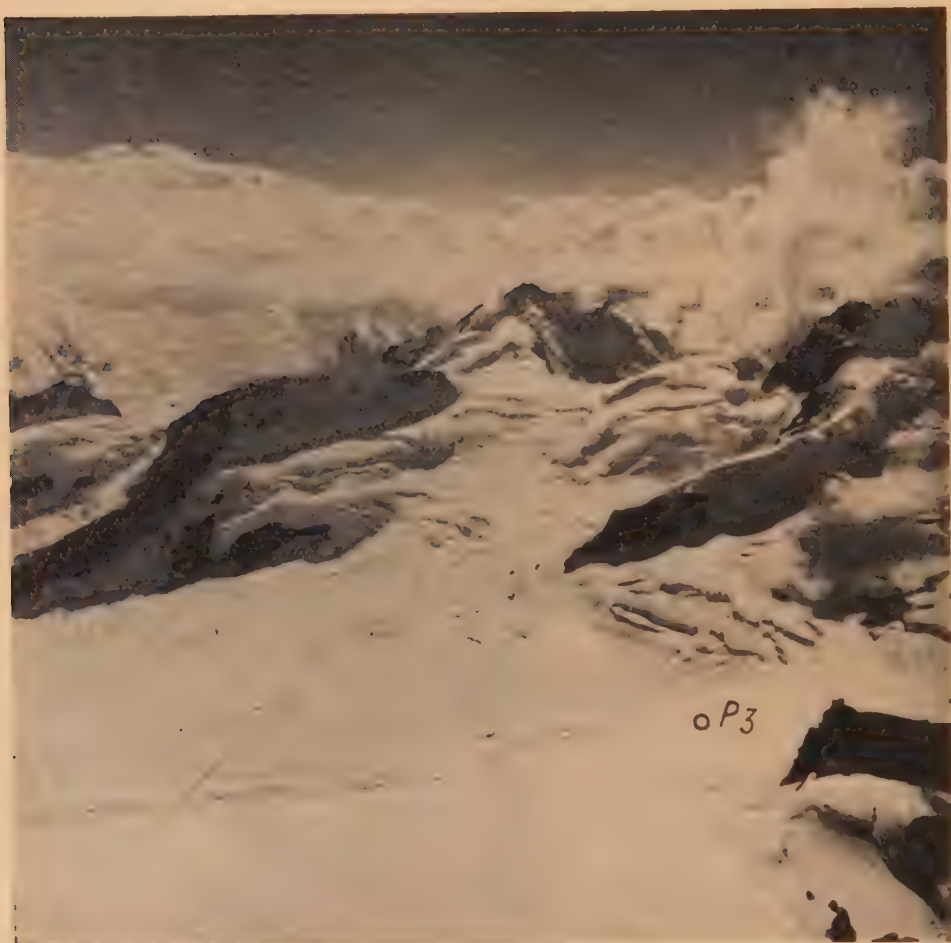


Fig. 2, Firn Region

- 2) Single measurement of snow and firn temperatures.
- 3) Observation and measurement of the motion conditions on the surface and below the surface of the glacier.
- 4) Observation and measurement of the alteration of firn and glacier surfaces, including ablation.
- 5) Disclosure of the topography of glacier bed by seismic means and under special consideration of its longitudinal profile.

In order to obtain a synthesis of natural observation, theory and experiment, the above mentioned operations 1-5 must be completed by respective theoretical and experimental investigations. In the meantime different laboratories in our country and abroad have started to clarify the plastic behaviour of ice to which the crystallographic investigations, executed at Jungfraujoch by the Geotechnic Commission have already contributed significant results [4, 5 and 6].

The uninterrupted continuation of operations, originally supported by the Swiss Commission for Snow and Avalanche Research and the Swiss Alpine club, is due to the Glacier Commission of the Swiss Society for the Advancement of Science (S.N.G.), who are the actual promoters of operations since 1944. Helpful suggestions have been received from the pioneer work of the Rhone Glacier Survey [7] and the investigations of Clariden Firn by the Glacier Commission of Zurich [8]. Energetic support is due to the Federal Topographic Service, the Laboratory for Hydraulic Research and Soil Mechanics of the Federal Institute of Technology, Bureau of Hydrology and Central Institute of Meteorology, Zurich [9]. To begin with, the execution of the programme mostly lay in the hands of A. Roch, whilst Mr. Pohl, Mr. A. Stavöstrand, A. Schönholzer, F. Denzler, H. Wiederkehr, A. Süstrunk, and R. Florin kindly contributed to the further operations, the latter two by seismic soundings. We gratefully accepted the hospitality offered by the board of directors of the Jungfraujoch Scientific Station.

II. Climate, Firn Augmentation and Firn Temperatures

For the performance of subsequent observations we had to be satisfied with both a relatively large-mesh network of observations and with single measures, the latter giving but a rudimentary view. It was renounced to accomplish measurement of radiation, condensation and evaporation.

1) Meteorological Data

Concerning climatic conditions, the Central Institute of Meteorology (M.Z.A.) has made available data of air temperatures and precipitation, which are represented graphically in Fig. 3.

Temperature fluctuations are shown in Table 1.

Table 1 : Air Temperatures at Jungfraujoch Station 1940-1947.

Extreme value	°C	Date
Lowest single Temperature	— 36.9	14 Febr. 1940
Highest single " "	+ 8.4	23 Aug. 1944
Lowest mean monthly Temperature	— 19.7	Jan. 1945
Highest mean montly " "	+ 1.7	Aug. 1945
Lowest summer mean " "	— 2.3	1941
Highest " " " "	— 0.4	1946
Lowest winter " "	— 12.2	1940/41 + 1941/42
Highest winter " "	— 10.0	1942/43
Lowest mean annual " "	— 8.9	1940/41
Highest " " " "	— 7.3	1942/43

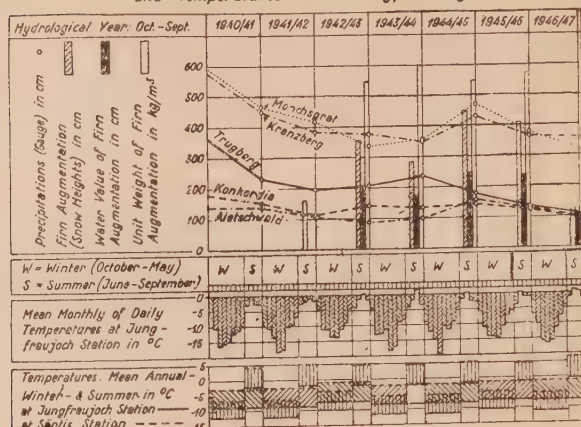
The winter season is considered as lasting from October to May and the summer season from June to September. The zone actually controlled being situated between 2.700 m (Concordia) and 4.000 m above sea level, the temperatures of the lower Sântis Station (2500) have been recorded (see Fig. 3).

The annual precipitation has been recorded by using 5 gauges belonging to the M.Z.A., the situation of which may be seen in Fig. 1. It still remains to be determined how far these precipitation recordings may be applied to our zone which covers a typical climatic shed.

2) Firn Augmentation and its Water Equivalent

The firn augmentation is determined at different spots at the end of a hydrological year by measuring the thickness of the snow at certain points, covering the snowlayer of the preceding summer. The thickness may be checked by means of snow depth gauges or by sounding down to the hardened summer snow or by compa-

Fig. 3 Precipitations, Firn Augmentation (near Point 3) and Temperatures in the Jungfrau Region



rative driving diagrams. For continual measuring, snow depth gauges or buoys are the best to rely on. In the Jungfrau zone they have been supplied by transoms which will enable a check by telescope up to a distance of 5 km. Thanks to the Army Avalanche Service it was possible to check simultaneously a number of 12 snow depth gauges during the war. In Fig. 4 the respective results of the good snow conditions in winter 1944/45 are represented in comparison with the lower stations of Davos and Weissfluhjoch. It is shown clearly that with increasing altitude, the maximum thickness of the snowlayer is reached later in the year. The maximum thickness of snow, over 9 m, at the highest point 5 is characteristic for the north-west margin of the observed zone, which is situated leeward of the main ridge. Point 3 may be considered as representative for a relatively large portion of the firn region (3350 m).

Schneehöhen in Davos, auf Weissfluhjoch und Jungfraujoch Winter 1944/45

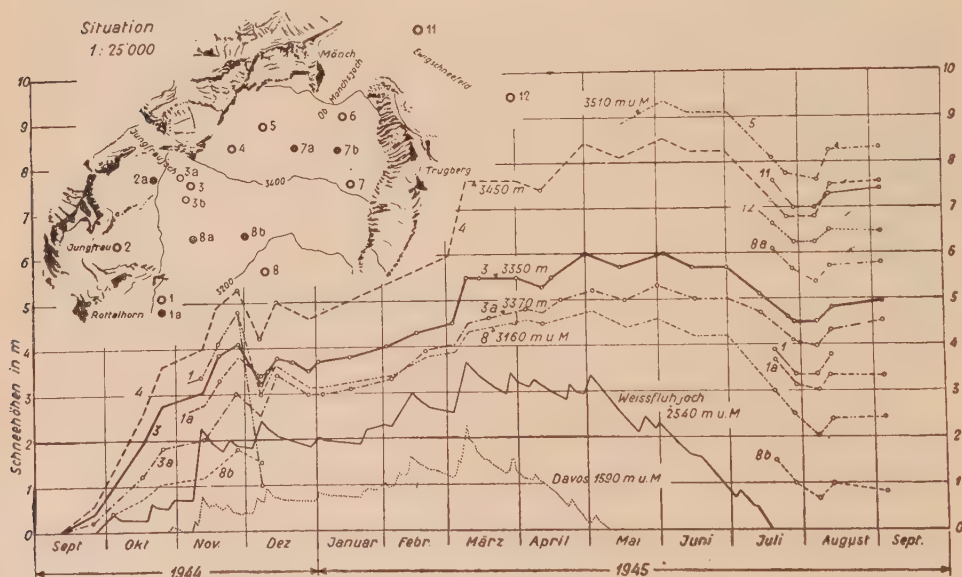


Fig. 4. Heights of Snow during the Hydrological Year 1944/45

The decrease of the thickness of the snowlayer during the short summer time (July and August) at the higher snow depth gauges is relatively small and scarcely exceeds 2m. The decrease is effected partly by ablation and partly by firnification of the snow and its consequent settlement, or respectively the compacting of the material. The firn augmentation in the higher firn regions (3400-4000 m) therefore largely depends on the quantity of snow and is influenced by ablation or by the mean summer temperature only to a relatively small degree, in contrast to what occurs in lower zones. The water equivalent is determined by measuring the unit weight. This operation which formerly required a tiresome excavation of the firn layer, has been facilitated by the construction of a firn boring apparatus, which allows the firn to be sampled down to depths of 7 m. Respective results are shown in the following table 2 and represented in Fig. 3.

Table 2 : Firn augmentation and its Water Equivalent of Snow Depth Gauge 3 (1942-1947).

Hydrological Year		42/43	43/44	44/45	45/46	46/47	Means
Firn Augmentation	cm	356	285	450	410	145	329
Unit Weight (Average)	kg/cm ³	548	600	548	575	773	584
Water Equivalent	cm	195	171	247	236	112	192
Annual Precipitation	cm	207	237	176	135	107	172
(Gauge Trugberg)							

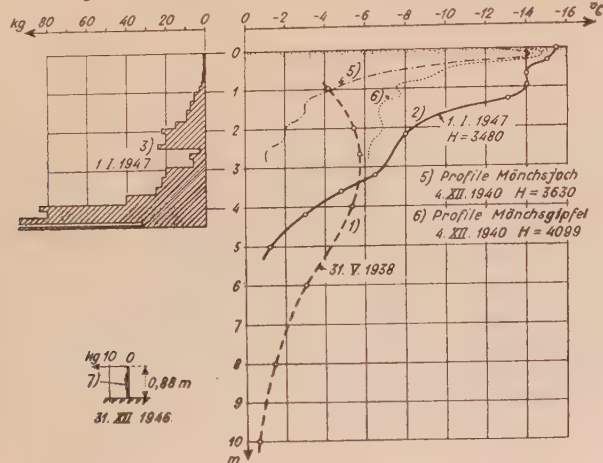
The noticeable increase of the unit weight with decreasing firn augmentation ist conditioned by the fact that at a certain ablation the firnification and glacification in the snowlayer increase with decreasing thickness of the layer. A noticeable decrease of the average unit weight with increasing altitude is consequently to be expected.

The increase of firn augmentation with increasing altitude is especially intens in the Jungfrau zone. Basing on snow depth gauge 3 e. g. at an approximate situation of the firn line at 3050 m. above sea level, a gradient of firn augmentation of 1.5 m. per 100 m. difference of altitude is to be determined for the hydrological year 1944/45. Due to the influence of the weather shed however, it is to be considered that the influence is not purely one of altitude.

3) Firn Temperatures

We contented ourselves with some complementary measurings of temperature during the high winter, reaching down to depths of 5 m only, as in 1938 Seligman and his assistants had made significant measurings of temperatures at the Jungfrau-

Fig. 5 Driving Diagrams and Firn Temperatures



firm. All of these show an extremely high gradient of temperature increase with the depth. The temperature curve of 1st January 1947(2) may be considered as a typical example and is shown in comparison with the temperature distribution recorded on 31st May 1938 by Seligman (1) in Fig 5. Considering the driving diagram (3), referring to curve (2) it is to be seen however, that the augmentation at the moment of recording has reached a thickness of about 4 m. and, compared with the snowlayer of Weissfluhjoch (2540 m) showed the remarkable compacting which is significant for high, windexposed firm zone (7).

In spite of the mean annual air temperature being of about -8°C , it is especially striking that the 0° -isotherm, and or the pressure melting point, seems to be situated at a relatively slight depth beneath the firm surface. G. Seligman in June determined a temperature of 0°C at a depth of only 15 m. The observation that the hanging glaciers situated at about 3400-3700 m. do not show a thickness exceeding 60 m. (see Fig. 2), allows the conclusion to be drawn that the 0° -isotherm, which under relatively low pressure may coincide with the line of the pressure melting point, is by no means situated deeper than 60 m, since a sufficiently fast sliding motion of the ice on the rock surface is only possible if the pressure melting point is reached at the contact surface. Due to the changes of air temperature, the 0° -isotherm too is subjected to certain annual fluctuations of which however we know nothing precise for the time being.

Single measurements carried out hitherto suggest that the 0° -isotherm during severe winter seasons is situated higher than during any later season, but this must first be proved by continual long-term measurings. The study of temperature conditions and of the heat balance of the firm cover has to include different influences: New snow, falling often at relatively high temperatures acts as a heat source [10]; the very small heat and temperatures conductivity of the snow cover in winter, the shift of phase involved by the penetration of cold waves [11], and especially the disturbing effect of the water from the snow melted in summer, which freezes when penetrating colder layers and consequently emits warmth.

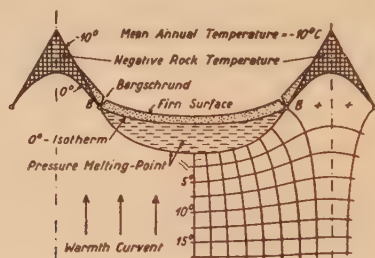


Fig.6 Temperature Conditions of a Firm Basin and its Subsoil

If the cross-section of a firm basin laterally confined by two sharp ridges is considered with reference to its rock bed, we may determine the general temperature distribution as represented in Fig. 6, based on former observations and reflections. The 0° -isotherm being located not far below the firm surface and supposed to be subjected to above-mentioned annual fluctuation, the rock surface is located in the zone of the pressure melting point. Both the rock surface and the 0° -isotherm confine the quasi-isothermic zone of the glacier ice and join laterally at the points B where the Bergschrund originates. Above Bergschrund the 0° -isotherm leaves the rock surface and reaches its maximum distance from the rock surface below its culmination. The rock and ice contact zone between the two points B, where the temperature is approximately 0° , must be intersected orthogonally by the stream lines of the current of earth warmth. Based on this condition we may determine the form of the heat flow as represented in Fig. 6, which allows the conclusion to be drawn that there is a certain concentration of the warmth current towards the firm basin.

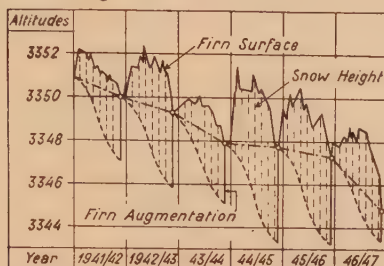
III. Height Variations of the Glacier Surface

The variations of the glacier surface have been checked partly at different points and partly in different sections. A more comprehensive investigation, based on photogrammetrical survey, is suggested to be carried out only towards the end of the 1940-1950 decade.

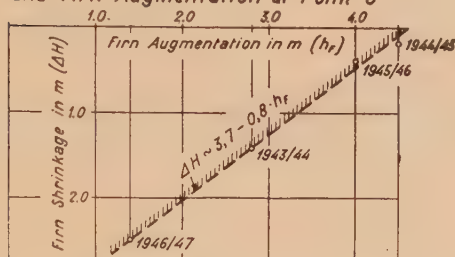
1) Continual Height Variations of Point 3 (1941-1947)

The height variations of an optically fixed point 3 (Fig. 1) on firn surface and on respective summer layers, measured periodically and represented in fig. 7a, show the following connections :

Fig. 7
a) Continual Height Variations of
Firn Surface, Snow Heights and
Firn Augmentation at Point 3



b) Relation between Firn Shrinkage
and Firn Augmentation at Point 3



a) The annual minimum height of firn is reached during formation of the summer layer after complete snow disappearance. These summer firn heights (chain dotted line) tended to incline steeply during the period 1941-1947. As to be expected and to be seen from Fig. 7b, the annual firn shrinkage increases with decreasing firn augmentation. For the maintenance of the firn height an average firn augmentation of at least 4.6 m would have been required, whilst the effective average amounted only to 3.2 m. It is also to be considered that a certain compensation exists by the fact that during a year — referring to broken line in Fig. 7a — the settlement of a certain summer layer decreases with increasing firn augmentation. The sine form of this curve shows that the respective layer sinks faster in winter than in summer. The total retreat of the firn height at point 3 amounted to 3.5 m from 1941 to 1946, representing a mean annual retreat of 0.7 m. The mere shrinkage during the record year 46/47 amounted to 2.5 m and therefore, represents more than three times the annual mean of 5 years.

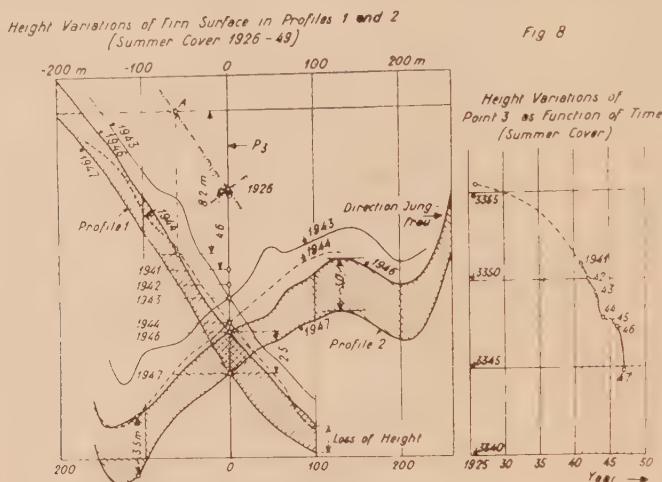
b) Due to annual lowering of the firn surface, the maximum firn height of the respective year appeared usually some months prior to the maximum thickness of the snow layer during this period. With increasing firn height this sequence would be reversed.

2) Annual Height Variations of the Firn Surface (1851-1947)

To determine the height variations of the firn surface of a large region, the observation of a single point is not sufficient. As completion of the measurements described in the foregoing paragraph, two profiles (1 & 2) have been placed through point 3, nearly at right angles to each other, and the annual variation of which have been recorded by levelling (a). By means of 3 transverse profiles the height variations of the glacier surface, which took place since 1926, have been determined (b) and two single points have been traced back to 1851 (c).

a) Annual Firn Shrinkage at Snow depth gauge 3 (1926-1947).

The variations of the two 300 m respectively 420 m long profiles intersecting in point 3 are represented in Fig. 8. This figure shows that the losses in height



stated at point 3 are not accidental but correspond to mean values. By means of profile 1 it is possible to trace the variations of a single point A of the firn surface back to the first photogrammetric surveys by the Federal Topographic Service in 1926. For point 3 the following losses of heights were obtained :

Table 3 : Firn Shrinkage at Snow depth gauge 3 (1926-1947).

Period	Number of years	Total Loss of Height in m	Mean annual Loss of Height in m
1926 - 1946	20	8.2	0.41
1926 - 1941	15	4.6	0.31
1941 - 1946	5	3.6	0.72
1946 - 1947	1	2.5	2.5

The extraordinary firn shrinkage of the hydrological year 1946/47 is represented in Fig. 8 by hatched area. The loss of height in profile 2 amounted to a maximum of 3.5 m.

b) Annual Firn Shrinkage in the Profiles Firn, Eck and Concordia. (1926-1947)

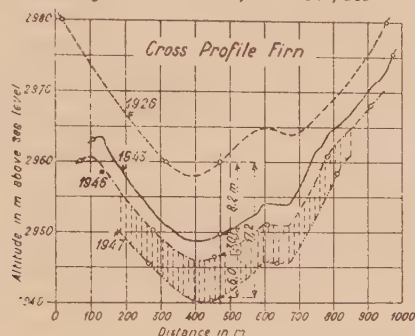
The variations of Profile Firn (see Fig. 1) in the 21 years period from 1926-1947 are shown in Fig. 9. Herewith the progressive character of glacier shrinkage during the past 2 decades is represented as at snow depth gauge 3. The comparison of long-term mean values, shown in Table 4 below, enables the conclusion to be drawn that usually the shrinkage of the ablation region exceeds that of the firn region and therefore increases with decreasing altitude, involving a small glacier surface inclination increase of some per thousands.

Table 4 : Mean Annual Shrinkage in m (1926-1947).

Period	Point 3	Profile Firn Snow d.g 9	Profile Eck Centre	Profile Concordia Centre
1926 - 1947	0.41	0.56	0.80	1.00
1946 - 1947	2.50	6.00	—	3.00

Note the extreme shrinkage of 6.00 m at snow gauge 9 (profile Firn) during the year 1946-47.

Fig.9 Height Variations of Firn Surface



c) Annual shrinkage at 2 points, observed since 1851 (1851-1947).

During the period of 96 years, points III and VIII, levelled in the Stingelin map were subjected to following losses of height :

Table 5 : Losses of Height at Points St. III and St. VIII in m (1851-1947).

Point	Altitude above Sea Level 1851 in m.	1851-1926		1926 - 1947		1851 - 1947	
		Total	Mean ann.	Total	Mean ann.	Total	Mean ann.
St. III	2993	15	0.20	15	0.71	30	0.31
St. VIII	2798	32	0.43	20	0.95	52	0.57

These values too show clearly both the glacier shrinkage increase with decreasing altitude and the very intense shrinkage during the past two decades, which in the region of Concordiaplatz amounted to an average of nearly 1 m. a year.

3) Consequences of Glacier Shrinkage

Among the numerous consequences of glacier shrinkage the following are to be mentioned :

a) Assuming an average glacier shrinkage of about 2.5 m over the whole Swiss Alpine region for the year 1946/47 the respective water value will exceed 100 % of the total annual precipitation. Considering the great potential energy of glacier water, it shows clearly that glaciers contribute remarkably to the power supply of the country, especially during dry and warm years. On the other hand it is to be considered that the actual abundance of water in the streams is based on a relatively fast removal of firn and ice reserves, accumulated during past decades.

b) Among dangers involved by glacier shrinkage, the glacier floods are to be mentioned, which consist in a sudden outbreak of huge quantities of water, stored in the interior, at the surface or at the sides of a glacier. If the glacier shrinkage were of long duration and of great amount the herewith involved relief might influence the balance of the earth crust [12]. The question arises if the present progressing uplift of large glacier regions as in Scandinavia and Greenland for instance, is really only caused by plastic after-effect of ice-melting since the glacial-era, or if an essential part of these uplifts may not be caused by elastic behaviour involved by progressive ice shrinkage during the past 100 years.

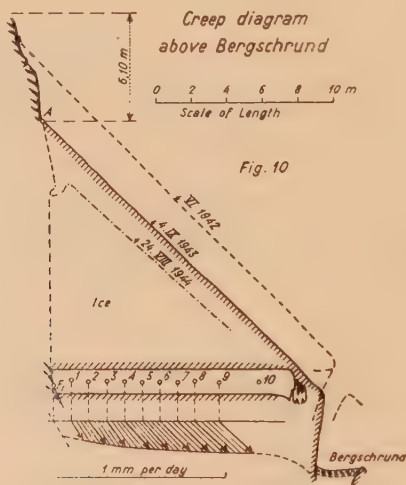
IV. Flow Velocities at the Surface and in the Interior of a Glacier

With the aid of subsequently described measurements the motion conditions above and below the Bergschrund, the formation of cracks, the annual fluctuation of the velocity and the effective shift of a firm grain should be accurately investigated, and preliminary data concerning velocity distribution along a vertikal be obtained (velocity diagram).

1) Creep Diagram above Bergschrund

Investigations have been carried out on the firm slope at the exit of the Sphinx-tunnel below Geneva Observatory, which on the downstream side is bounded by the Bergschrund and shown in longitudinal section in Fig. 10. Snow depth gauge measurements during the first winter season 1941/42 proved the quantities of snow deposited above Bergschrund to be merely a very small part of the snow quantities below Bergschrund. This upside snow deposit is carried into and across the Bergschrund by wind action, by continual flow on the slope surface and by avalanche action. Under these conditions the firm and ice cover can only be maintained at this place if the ice motion is slower in relation to the smaller firm augmentation above the Bergschrund than below it. This is the reason for the fast opening of Bergschrunds during periods of little precipitations.

Therefore the common definition of Bergschrund as the separation between the solid and mobile glacier portions is only approximately correct. Accurate observation proves the ice mass to be subjected to a slow creep motion above Bergschrund. Consequently the upper lip advances slowly, involving a new fissure higher up, until the ice mass between the two cracks falls into the old Bergschrund or passes over it.



For the purpose of more accurate investigation of motion conditions above Bergschrund a horizontal gallery was dug in summer 1941, allowing access to the rock surface. It has been provided with measuring bolts, the velocity components of which have been recorded relatively to the rock bench mark F1 during several years. In Fig. 10 the mean velocities of the wooden bolts in the ice are represented as a horizontal creep diagram for the period from 1st April 1942 to 26th August 1944. The parabolic course of the diagram corresponds to an approximately constant toughness of the ice.

At the innermost point, situated about 1 m from the rock, the resulting velocity amounted to 0.14 mm per day; at the outermost point 9 at an approximate distance of 9 m from rock, however, it amounted to about twice as much. In spite of the partly lower temperature, a progressive velocity increase was ascertained in the superficial firm cover in accordance with the lesser viscosity (near point 10). Since it could be observed how a rock slab had been torn away from the parent rock and carried along with the ice, and further that the contact zone of the ice is inter-

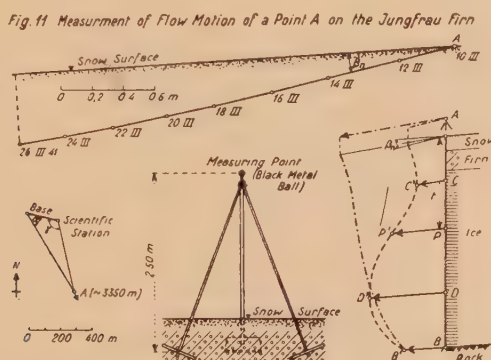
persed with broken pieces of rock, it must be assumed that a very small sudden increase in velocity probably occurred in the boundary surface between rock and ice. Consequently, the glacier above Bergschrund too has an erosive power not to be underrated. For comparison we further mention the short-term observation of a point situated on the under-lip of Bergschrund, the shift of which amounted to 8 cm a day, representing a value 300 times greater than the furthest extremity of measuring point 9 of the upper lip.

2) Motion Conditions below Bergschrund

In the following, the motion conditions of a region extending to below the firn line are described :

a) Effective Velocities of single Points of the Firn Surface and their Annual Fluctuations

In order to recognise the connections between winter snow cover creeping and glacier motion, we tried to determine accurately the local motion of a point on the firn surface for the first time in winter 1940/1941 (near depth gauge 3). Carried on a tripod by an already compacted snowlayer situated somewhat beneath the surface, a black metal ball acted as measuring point. From base I, established on the Jungfrauoch, the shift of the ball was recorded trigonometrically during a fortnight. The evaluation of these measurements led to the following result represented in Fig. 11, the mean daily velocity component parallel to the snow surface amounted



to 17.2 cm, the perpendicular component to 2.7 cm and therefore the mean daily creep-angle β_n to about $9^\circ 13'$. Later measurements proved this creep-angle also to be subjected to annual fluctuations and generally to be smaller in midsummer than in midwinter. If this measurement is carried out, during a whole year, the mean creep-angle becomes identical with the dive-angle Φ introduced by Finsterwalder. We can then state :

$$1) \quad \tan \beta_n = \tan \Phi = \frac{v_y}{v_x} = \frac{A}{v_x}$$

where : v_x = annual shift parallel to surface

v_y = annual shift perpendicular to surface

A = thickness of firn augmentation

Measurements of the angle Φ at different points situated between the firn line and point 3, confirmed the connection represented by equation 1). With an observed annual 46 m shift of point 3 and a 4.6 m firn augmentation of the stationary glacier, we obtain $\tan \Phi = 0.10$, i.e. $\Phi = \text{abt. } 6^\circ$.

The systematic observations required to determine the annual fluctuations of surface velocity, are still very incomplete. The best measurement to rely on dates from 1941, resulting in a mean daily velocity of 15.4 cm in March, 11.3 cm at the beginning of May and 7.4 cm at the end of August and was recorded at a point near snow gauge 3, showing that the probable maximum velocity of February was more than double the summer minimum value. In later years, however, measurements showed only a small difference between summer and winter velocities. Yet

— contrary to the ablation region — they also mostly confirm the fact of the firn region surface flowing faster in winter than in summer, which has already been determined at Hintereisferner by A. Blümke and S. Finsterwalder [14] and at Clarden firn by R. Streiff-Becker [8]. This striking phenomenon may be caused by the change of ice toughness due to the higher winter loading (see Fig. 7), the influence of the annual temperature fluctuation in the firn cover and the annual fluctuation to be expected in crack formation and their corresponding contingencies.

b) Velocity Distribution of Glacier Surface in Conjunction with Crack Formation

A series of surface velocities, or their horizontal components, measured at different superficial points and at two profiles Firn and Eck, are represented in Fig. 12 and show seasonal and annual values. With the exception of the relatively great velocities of two points at 3828 m and 3740 m above sea level, situated beneath Rottalsattel, a distinct flow velocity increase from 8 to 15 cm a day was observed between Bergschrund and firnline. Below the firnline the damming ac-

Fig. 12 Surface Velocities



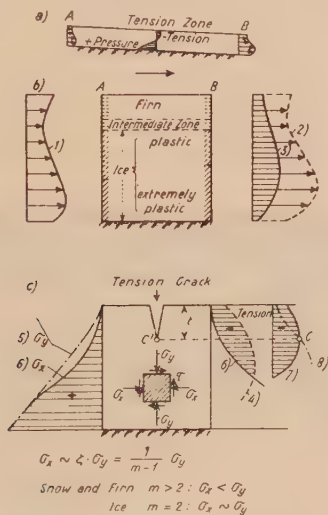
tion of Concordiaplatz may already be noticed through a general slow down rendered externally visible by the non-appearance of crevasses. Accelerated motion causes tension, whilst delayed motion causes pressure stresses. Similar to the snow-cover, it is therefore possible to distinguish between tension compression and neutral zones also in a glacier. Leaving the most-effective bending and shear stresses out of consideration, the formation of tension and cracks in the tension zone according to graph in Fig. 3 may be explained as follows :

The specific elongation of the tension zone of a glacier situated between two cross-sections A and B, coincides with the difference of the two velocity diagrams 1 and 2, marked by hatchings (3) in fig. 13b. The elongations are caused by tensions, the diagram of which is represented by curve 4 in Fig. 13b. These tensions are superposed on horizontal gravity stresses σ_x , the distribution of which is represented by curve 6. The difference of the two diagrams 4 and 6 therefore results in the hatched diagram of the tensions 7, which are effective in the upper layers of the section. Completing this drawing by the tensile strength of the respective layers (8), it is obvious that the maximum depth of the crevasses is represented by point C, where both curves 7 and 8 intersect. For the purpose of comparison we mention that on the Jungfraufirn no crevasses with depths of more than 30 m were discovered.

Besides that, the crack formation process is characterised by the tensile fields being mainly given by the morphology of the subsoil and the glacier section, i.e. being stationary, whilst the glacier ice flows across the tensile fields. Observing the

appearance and disappearance of a crevasse, it becomes obvious, that it appears at a certain position A, reaches its maximum dimensions at B and disappears again at C, whilst at A in the meantime a new crevasse is forming. A crevasse of 18 m depth and 35 cm opening, for instance, was observed and showed an opening velocity of 8 mm a day. Since this crevasse is situated not far below point 3, where the specific elongation velocity of the surface firn layers may be estimated at 0.2 ‰ a day on the basis of the velocity measurements, the above mentioned opening velocity would completely adjust the differences of flow velocities existing within a distance of 40 m.

Fig.13 Formation of Tension Cracks



c) Velocity Distribution in the Cross-Section of the Glacier (Profile Concordia)

For several years projects have been mooted but not yet realized, for the measurement of velocity distribution along a vertical line by sinking a system of flexible metal tubes into a bore hole and checking its inclination changes with the help of an electric clinometer. We nevertheless contribute a reply to the main questions as to whether a glacier flow velocity is greater at the surface or in the interior. At Concordia profile we determined the average flow velocity of the glacier surface and we tried to estimate the mean annual velocity v_E of the glacier, in relation to the whole cross-section based on the quantity balance. Here the known formula is valid :

$$2) \quad v_E = \frac{Q_E}{F}$$

where Q_E signifies the quantity of ice flowing annually through the cross-section F. Whilst the cross-sectional area F has been determinated by seismic soundings, the following formula of the annual balance was used to determine the quantity of ice Q_E :

$$3) \quad Q_E = (E_F - E_a) - \Delta F + a$$

where :

- E_F = volume of firn augmentation translated to ice
- E_a = volume of ablation in the region above the cross-section
- $\Delta F + a$ = change of glacier volume above the cross-section
 - Augmentation or accumulation = positive
 - Decrease or shrinkage = negative

In equation 3 the shrinkage of the ice, caused by melting process in the inte-

rior and at the sole of the glacier is neglected. The insignificance of this shrinkage is made obvious by the winter discharge, which amounts to about 1.9 lt./sec. km², representing an average ice thickness of 6 cm. a year in the whole glacified basin above Concordia profile. This value represents a limited maximum value as any possible springwaters are included.

Assuming an average 2 m. firn augmentation, a 1.5 m. ablation and an annual shrinkage average of 1 m., we obtain from equation 3 :

$$Q_E = (2.53 - 1.5 \cdot 22 + 1.0 \cdot 75) 10^6 = 148.10^6 \text{ m}^3/\text{year}.$$

Under reservation of a definitive evaluation of the seismic soundings carried out by A. Süssstrunk and R. Florin in summer 1947, we assume the maximum depth of the cross-section not to exceed 500 m. For a parabolic cross-section 500 m deep and 1500 m broad, an average transit flow velocity of 300 m/year may be calculated with aid of equation 2.

On the other hand the maximum surface velocity was determined to be 205 m in 1945/46 and 207 m in 1946/47, from which we can calculate a maximum annual velocity of 206 m, representing a value which is only about 3% smaller than the mean summer velocity of the same point, measured always in September (see Fig.14).



Fig. 14. Concordiaplatz (from Kamm)

Based on the measurements at 7 points distributed over Concordia profile, the mean surface velocity of about 80 % of the value, i.e. 165 m/year, may be assumed. Therefore it follows, that the mean flow velocity v_E the preliminary value of which was calculated as 300 m/year, is considerably greater than the average surface velocity. Considering furthermore that the maximum flow velocity of the whole cross-section must be considerably greater than the average velocity v_E at the cross-section, it becomes obvious that at a certain depth the glacier must flow faster than at the surface. Herewith the existence of a maximum velocity at lower depths is fundamentally confirmed by the above-mentioned investigations which still require a later precise interpretation. This has already been recognised by A. Heim [15], R. Koechlin [16], R. Streiff-Becker [8] and M. Demorest [17] and may also be derived from plastic behaviour of ice [18]. It is to be considered, however that the observed cross-section is already situated in the ablation region of the glacier. Since, on the other hand an increase in the creep velocity with the depth was observed in the upper surfaces of the Jungfraufirn, analogous to that in a layer of winter snow, it is to be assumed that in the firn region a minimum exists in addition to the maximum (cf. Fig. 12).

3) Various Observations in the Ablation Region

According to a guide's statement, at Easter 1919 a recently fallen block, as high as a house, was situated in the glacier surface at the southeast foot of the Kranzberg (near point 2799 m). This block (called Concordiablock) was measured by the federal Topographic Service in autumn 1926. In September 1946 it was at about the level of the «second» triangle at the foot of Klein-Dreieckhorn; it had travelled a distance of about 3.8 km. in 20 years, which represents an average velocity of 190 m. a year.

A clear idea of the relatively great surface velocity in the zone passed by Concordiablock is given by the patches on the glaciers's axis, visible in Fig. 14. They originate in a discontinuous moraine formation, which is to be attributed to the rock slide at point 2902 m; this slide usually reaches its maximum intensity at the moment of complete disappearance of snow. From a bird's eye view all these patches resemble a broken line running between the two central moraines, which bound the glacier originating from the Jungfraufirn. The distance between two such patches represents the annual velocity of the glacier surface at the respective position.

Whilst at profile Eck the Jungfraufirn surface is 1600 m wide, this decreases to a tenth, namely 160 m., at Concordia profile after a distance of about 2 km. This is made obvious by the course of the respective central moraines (Fig. 1 and 14). The effect of the glacier being thus crushed together causes the ice to form a distinct slaty structure. Jungfraufirn is consequently pushed away, except for a narrow strip of the glacier surface, by its mightier neighbours, Great Aletsch Glacier to the right and Ewigschneefeld to the left, and this obviously causes the main quantity to dive, so to speak, at the entrance into the Concordiaplatz.

From this it also follows that at profile Eck the longitudinal section of the glacier sole might show a steep step, which corresponds to the great ice depth of the Concordiaplatz, determined by echo-soundings [2] to be a maximum of 790 m. On the other hand a recess in the rock sole in the centre of the Concordiaplatz was ascertained in 1947, and in it the ice mass of Jungfraufirn is embedded. Such a dive favours the grinding of deep glacier beds, the formation of which is observable mainly at the confluence of several glaciers. The formation of rock shoulders might also be explained by this phenomenon. Furthermore if the laws of formation of «roches moutonnées» [19] are applied to the general erosive action of a glacier, we conclude that at the exit of Concordiaplatz a rock threshold may be expected, quite independent of the geological properties of the subsoil. Its formation may be explained by decreasing erosive power at the luff-side of an obstruction. This decrease of erosive power is conditioned by increasing plasticity of the ice within zones of high pressures, as expected for instance at the discharge across the narrow Concordia section [5], [20]. The existence of such a rock threshold beneath Concordiaplatz has been confirmed by echo-soundings made in summer 1947.

BIBLIOGRAPHY

- [1] O. LÜTSCHG : Der Märjensee und seine Abflussverhältnisse Annalen der Schweiz. Landeshydrographie, Bern 1915.
- [2] H. MOTHES : Neue Ergebnisse der Eisseismik, Zeitschrift für Geophysik, Jg. 5, H 3/4, Braunschweig 1929.
- [3] G. SELIGMAN : Forschungsergebnisse am grossen Aletschgletscher, Die Alpen, H. 12, 1943.
- [4] R. U. WINTERHALTER : Probleme der Gletscherforschung, Die Alpen, 1944, H. 6.
- [5] R. HAEFELI : Entwicklung der Probleme der Schnee- und Gletscherkunde in der Schweiz, Experientia 2, H. 1, 1946.
- [6] M. DE QUERVAIN : Schnee als kristallines Aggregat, Experientia 1, S. 207, 1945.
- [7] P. L. MERCANTON : Vermessungen am Rhonegletscher 1874-1915, Neue Denkschrift der Schweizerischen Naturf. Ges. 1918
- [8] R. STREIFF-BECKER : Beitrag zur Gletscherkunde, Forschungen am Claridenfirn im Kanton Glarus, Denkschrift der Schweiz. Naturf. Ges., Bd. LXXV, 1943.
- [9] Annalen der Schweiz. Meteorolog. Zentralanstalt (MZA) in Zurich, bis incl. 1946.
- [10] M. DE QUERVAIN : Zum Wasserhaushalt der Schneedecke ; Hydrologische Tagung der Eidg. Kommission für Schnee- und Lawinenforschung ETH 1947. (interner Bericht).
- [11] E. SORGE : Glaciologische Untersuchungen in Eismitte. Wissenschaftliche ergebnisse der deutschen Grönlandexpedition von K. Wegener, Bd. III, Glaciologie, Leipzig 1935.
- [12] B. GUTENBERG : Changes in sea level, postglacial uplift and mobility of the earth's interior, Bulletin of Glaciological Society of America 1941.
- [13] R. HAEFELI : Spannungs- und Plastizitätserscheinungen in der Schneedecke Mitteilung Nr. 2 der Versuchsanstalt für Wasserbau an der ETH, 1942.
- [14] A. BLÜMKE und S. FINSTERWALDER : Zeitliche Aenderungen in der Geschwindigkeit der Gletscherbewegung, Sitzungsbericht bayer. Akad. Wiss. Kl. Bd. 35, 1905.
- [15] ALB. HEIM : Handbuch der Gletscherkunde, Stuttgart 1885.
- [16] R. KOEHLIN : Mécanisme de l'eau et principes généraux pour l'établissement d'usines hydro-électrique, Kap. 11, Bd. 1, 1924/26, Les glaciers et leur mécanisme Lausanne 1944.
- [17] M. DEMOREST : Ice Flowage as revealed by glacial striae, Journal of Geography, Vol. XLXI, No. 5, 1938.
- [18] R. HAEFELI : Zur Mechanik ausserordentlicher Gletscherschwankungen Schweiz. Bauzeitung, Bd. 115, Nr. 16, 1940.
- [19] A. CAROL : Beobachtungen zur Entstehung der Rundhöcker, Die Alpen Nr. 6, 1943.
- [20] M. F. PERUTZ : Report on problems relating to the flow of glaciers, The Journal of Glaciology, Vol. I, No. 2, 1947.

5) DIE KURZFRISTIGEN SCHWANKUNGEN DER ALPENGLETSCHER UND IHRE URSACHEN

von

Dr. R. BILLWILLER,

a. Dir. der Schweiz. Meteorologischen Zentralanstalt und Präsident der Zürich. Gletscherkommission.

Die Nachrichten aus früheren Jahrhunderten sind natürlich spärlich und ungenau. RICHTER (1) hat alle aus dem ganzen Alpengebiete vorliegenden Angaben über den Stand der Gletscher zusammengestellt und fand, dass die Zunahme und Abnahme derselben in einer Periode erfolgt, die zwischen 20 und 45 Jahren schwankt und im Mittel der drei letzten Jahrhunderte 35 Jahre betrug. Das war auch die Länge der gerade damals von BRÜCKNER aufgestellten Klimaperiode und Brückner sah darin eine willkommene Stütze seiner Theorie. Seither wird die Realität dieser Klimaperiode allerdings stark angefochten.

Sichere Nachrichten über den Stand der Alpengletscher besitzen wir erst aus der Zeit nach 1800. Es darf immerhin gesagt werden, dass diese die Existenz einer tertio-säkulären Periode der Gletscherschwankungen zu bestätigen scheinen: Sehr ausgesprochene Hochstände um die Jahre 1820 und 1855, sodann in den Westalpen wenigstens ein Stillstand in deren Rückgang um 1890 und ein deutlich erkennbares Wachsen der Gletscher zwischen 1910 und 1925. Die beiden letztgenannten Phasen verwischen aber keineswegs den allgemeinen starken Rückgang nach 1860.

Unsere im Jahre 1913 gegründete «Zürcher Gletscherkommission» betrachtete es als ihre Hauptaufgabe durch exakte Bestimmungen des jährlichen Firnzuwachses oberhalb der Schneegrenze eine dauernde Ueberwachung der Sammelgebiete einiger Gletscher einzurichten. Sie wählte hierfür den Claridengletscher in den Glarner- und den Silvrettagletscher in den Graubündner-Alpen. Sie sorgte auch für die Ermittlung der in beiden Gebieten anfallenden Jahresmengen des Niederschlags durch Aufstellung von Totalisatoren. Da wir anderseits durch die zuverlässigen und homogenen Beobachtungen des Säntis über die Temperaturverhältnisse der höheren Regionen des Alpengebietes unterrichtet sind, so gewinnen wir erstmals für den letzten Vorstoss der Alpengletscher, der sich über die Jahre 1910 bis 1926 erstreckte, einen Einblick in die meteorologischen Bedingungen, unter welchen die Gletscher wachsen.

Wir geben im folgenden für den Zeitraum 1915 bis 1947 den Betrag des jährlichen Firnzuwachses, ermittelt jeweils am Ende des hydrographischen Jahres (September) für die obere Claridenboje in 2910 m, deren Exposition uns besonders günstig erscheint; ferner die im Totalisator des Gebietes im vorausgegangenen Jahre (Oktober-September) gemessene Jahressumme des Niederschlags und das Temperaturmittel des Säntis während der vergangenen Ablationsperiode (Mai bis September) samt dessen Abweichung vom Normale.

Der Beginn der Messungen der Zürcher Gletscherkommission fällt mitten in die letzte Wachstumsperiode und lässt daher keinen Vergleich mit den vorangegangenen Jahren des Schwindens zu. Dagegen erlauben die seit 1883 vorliegenden Temperaturmessungen des Säntis einen solchen wenigstens für die Temperatur. Wir stellen sie daher voran und fügen fortlaufende Fünfjahresmittel dieser Temperaturen bei, in denen die Summationswirkung einer Reihe von Jahren zum Ausdruck kommt. Das jedem Jahre beigegebene Jahrfünftmittel umfasst jeweils das mit dem betreffenden Jahre endigende Jahrfünft.

Aus dieser Zusammenstellung ist folgendes ersichtlich. Für die jährlichen Niederschlagsmengen lässt sich während der Wachstumsperiode der Gletscher von 1910-1925 und für die darauf folgende, bis heute andauernde Schwundperiode keine systematische Differenz erkennen. Dagegen zeigen die Sommertemperaturen des Säntis von 1909-1926 sehr nennenswerte negative, von dann an positive Anomalien. Negative Anomalien lassen sich in kleinerem Ausmass auch um 1890 feststellen, also für dieselbe Zeit, für welche sich in den Westalpen ein Stillstand im Schwunde der Gletscher erkennen liess. Das Wachsen der Alpengletscher fällt demnach zusammen mit trüben und kühlen Sommermonaten, in welchen im Firngebiet der Anteil des als

(1) Richter E.: Geschichte der Schwankungen der Alpengletscher, Zeitschr. des D. Oe. Alpenvereins, 1891

Jahr	Sonnentemperatur (VI-IX) des Sants	Abweichung	Jahresmittel	Abweichung	Firnzuwachs Clariden: (2910 m.)		Niederschlag dieser
					Periode:	Betrag:	Periode:
1883	3,2	—0,6	—	—			
1884	3,8	0,0	—	—			
1885	4,9	1,1	—	—			
1886	4,6	0,8	—	—			
1887	4,9	1,1	4,3	0,5			
1888	3,6	—0,2	4,4	0,6			
1889	3,4	—0,4	4,3	0,5			
1890	2,8	—1,0	3,9	0,1			
1891	3,4	—0,4	3,6	—0,2			
1892	4,4	0,6	3,5	—0,3			
1893	4,0	0,2	3,6	—0,2			
1894	3,1	—0,7	3,5	—0,3			
1895	4,9	1,1	4,0	0,2			
1896	2,8	—1,0	3,8	0,0			
1897	4,2	0,4	3,8	0,0			
1898	4,2	0,4	3,8	0,0			
1899	3,7	—0,1	4,0	0,2			
1900	4,9	1,1	4,0	0,2			
1901	4,2	0,4	4,2	0,4			
1902	3,5	—0,3	4,1	0,3			
1903	3,4	—0,4	3,9	0,1			
1904	4,4	0,6	4,1	0,3			
1905	4,9	1,1	4,1	0,3			
1906	3,6	—0,2	4,0	0,2			
1907	3,9	0,1	4,0	0,2			
1908	3,7	—0,1	4,1	0,3			
1909	2,4	—1,4	3,7	—0,1			
1910	2,6	—1,2	3,2	—0,6			

Schnee fallenden Niederschlags am Gesamtniederschlag vermehrt und die Ablation stark vermindert ist.

A. PENCK und A. BRÜCKNER (1) haben die ein ungleich grösseres Ausmass erreichenden Vorstösse der Alpengletscher in den diluvialen Eiszeiten als Perioden einer allgemeinen Temperaturniedrigung und nicht als Zeiten vermehrten Niederschlags angesehen. Was wir aus dem letzten Gletschervorstoss folgern müssen bestätigt diese Annahme.

Es wird die Aufgabe unserer Nachkommen sein, das Verhalten der meteorologischen Elemente während zukünftiger Gletscherschwankungen zu kontrollieren. Das ist zufolge vermehrter Beobachtungsmöglichkeiten — wir denken dabei in erster Linie an die Beobachtungen des Observatoriums auf Jungfraujoch — leichter als in der Vergangenheit. So werden sich Anhaltspunkte gewinnen lassen über die quantitative Auswirkung auf den Gletscherstand und Rückschlüsse auf die für die diluvialen Eiszeiten notwendigen Bedingungen. Einen ersten Versuch habe ich bereits vor Jahren gemacht (2) und dabei schon die Erkenntnis gewonnen, dass man bei Abschätzung der Wirkungen der meteorologischen Elemente auf eine Gletscherschwankung nicht auf die absoluten Gletscherstände abstellen darf. Der während einer Vorstossperiode erreichte Stand ist weitgehend bedingt durch den Stand der Gletscher bei dessen Beginn. So wird ein kommender Vorstoss auch bei für ihn sehr günstigen meteorologischen Bedingungen unsere Alpengletscher nicht auf den Stand um 1860 und 1820 zurückführen können. Denn die Alpengletscher sind gegenwärtig durch eine seither fast ununterbrochene Schwundperiode ganz ausserordentlich reduziert und werden ihre frühere Ausdehnung nur als Resultat mehrerer, jeweilen mit einem Grössenzuwachs abschliessenden Schwankungen erreichen können.

(1) A. Penck und E. Brückner ; Die Alpen im Eiszeitalter, Leipzig 1909.

(2) R. Billwiller : Temperatur und Niederschlag im schweizerischen Alpengebiet während des letzten Gletschervorstosses und einige Rückschlüsse auf die eiszeitlichen Verhältnisse. Annalen der Schweiz. Meteorolog. Zentralanstalt, Jahrgang 1930.

DIE KURZFRISTIGEN SCHWANKUNGEN DER ALPENGLETSCHER UND IHRE URSACHEN

von **R. BILLWILLER**, Zürich
(Resumé)

Aus dem aus dem ganzen Alpengebiete vorliegenden historischen Nachrichtenmaterial haben verschiedene Forscher (FOREL, RICHTER und BRÜCKNER) die Existenz von Schwankungen der Gletscher abgeleitet, deren Periode ungefähr $1/3$ Jahrhundert beträgt. Sehr ausgesprochen waren die Hochstände um 1820 und 1855; derjenige von 1890 zeigte sich nur andeutungsweise in den Westalpen und war mehr ein Stillstand im Schwinden, während zwischen 1910 und 1925 wieder ein deutlich erkennbares Wachsen sich ergab. Seither gingen die Gletscher stark zurück und die beiden letztgenannten Wachstumsphasen verwischen keineswegs den allgemeinen starken Rückgang nach 1860.

Die Zürcher Gletscherkommission bringt seit 1913 exakte Bestimmungen des am Ende jedes Sommers sich ergebenden *Jahresfirnzuwachses* bei. Der Schreibende bemüht sich auch aus dem bis jetzt vorliegenden meteorologischen Beobachtungsmaterial die Wirksamkeit der den Gletscherstand bedingenden Faktoren herauszuschälen.

Aus den bis 1883 zurückreichenden, homogenen *Temperaturbeobachtungen* des *Säntis* lässt sich mit Sicherheit nachweisen, dass es die *Sommertemperatur* und nicht etwa vermehrte Niederschläge sind, welche das *Wachstum der Gletscher* zwischen 1910 und 1926 bedingten. Trübe und kühle Sommermonate vermehren den Anteil des als Schnee fallenden Niederschlags und vermindern die Ablation.

Jahr	Jahrfünftel	Abweichung	Jahr	Jahrfünftel	Abweichung	Jahr	Jahrfünftel	Abweichung
1883	—	—	1908	4 ^o ,1	0 ^o ,3	1933	4 ^o ,5	0 ^o ,7
1884	—	—	1909	3,7	— 0,1	1934	4,4	0,6
1885	—	—	1910	3,2	— 0,6	1935	4,5	0,7
1886	—	—	1911	3,6	— 0,2	1936	4,7	0,9
1887	4 ^o ,3	0 ^o ,5	1912	3,2	— 0,6	1937	4,4	0,6
1888	4,4	0,6	1913	2,9	— 0,9	1938	4,5	0,7
1889	4,3	0,5	1914	3,0	— 0,8	1939	4,5	0,7
1890	3,9	0,1	1915	3,1	— 0,7	1940	4,0	0,2
1891	3,6	— 0,2	1916	2,4	— 1,4	1941	4,1	0,3
1892	3,5	— 0,3	1917	3,2	— 0,6	1942	4,3	0,5
1893	3,6	— 0,2	1918	3,3	— 0,5	1943	4,3	0,5
1894	3,5	— 0,3	1919	3,6	— 0,2	1944	4,4	0,6
1895	4,0	0,2	1920	3,6	— 0,2	1945	4,8	1,0
1896	3,8	0,0	1921	4,3	0,5	1946	4,9	1,1
1897	3,8	0,0	1922	3,9	0,1	1947	5,2	1,4
1898	3,8	0,0	1923	4,0	0,2			
1899	4,0	0,2	1924	3,8	0,0			
1900	4,0	0,2	1925	3,7	— 0,1			
1901	4,2	0,4	1926	3,4	— 0,4			
1902	4,1	0,3	1927	3,6	— 0,2			
1903	3,9	0,1	1928	4,0	0,2			
1904	4,1	0,3	1929	4,5	0,7			
1905	4,1	0,3	1930	4,8	1,0			
1906	4,0	0,2	1931	4,6	0,8			
1907	4,0	0,2	1932	4,9	1,1			

Dieses Resultat liegt ganz in der Richtung der von *Penk* und *Brückner* als Ursache für die diluvialen Eiszeiten vermuteten Temperaturenniedrigung ohne eine wesentliche Vermehrung des Gesamtniederschlags. (Vergleiche darüber R. Billwiller, « Temperatur und Niederschlag im schweizerischen Alpengebiet während des letzten Gletschervorstosses und einige Rückschlüsse auf die eiszeitlichen Verhältnisse ». Annalen der Schweizerischen Meteorologischen Zentralanstalt, Jahrgang 1930).

6) HORIZONTAL AGE DIFFERENCES IN GLACIERS

by

George J. HEINSHEIMER

During snow fall the new layers of snow become spread over the older ones. Its uniform distribution may be changed to some extent by the wind; but the influence of the wind is limited as to time: the pressure of the wind itself; melting and refreezing and the condensation of the snow after its deposition cause its surface to harden, and as a consequence less and less snow can be transported by the wind. The action of the wind then becomes restricted to the distribution of dust particles. When during the summer the snow surface melts due to air currents and rain, the melt water generally penetrates into the snow mass until reaching a frozen impervious layer, along which it flows out. But the dust remains on the surface and becomes more and more concentrated and ultimately forms the so-called dirt bands, well known to every glacier visitor. They appear more or less distinctly in every vertical section of the ice or névé, for instance in crevasses, and look very much like the annual rings of growth in wood, except that they extend horizontally in a series. Nevertheless, the age indication is not quite reliable. During hot and dry summers the whole deposit of a year or more may waste away, so that two or more dirt bands become united. An example of this was demonstrated by R. Streiff-Becker (Zwanzig Jahre Firneobachtung, Zeitsch. für Gletschk. 1936, p. 31, ff) by means of artificial dyes.

But when an ice granule once becomes embedded in the glacier deeply enough below the surface to be protected from melting, it may be added to another due to pressure and glacier movement, thus forming a bigger crystal; but its course is now mathematically fixed, as S. Finsterwalder has shown in his mathematical theory of glacier movement. To every point of a glacier above the névé line corresponds another below, both connected by stream lines.

All these facts are well known; but, now we shall go one step further: if the average velocity of an ice granule on its way from its entrance into the glacier to the point where it melts, be denoted by C_n then evidently the equation holds true:

$$y_n = c_n \cdot t_n$$

where S_n denotes the distance traversed in the time t_n .

But we know that the velocity of a glacier's movement, which depends on the season, the inclination of the ground, the thickness of the ice, etc., is always much greater in the central portion of a glacier than at its margins. Therefore, a granule situated in the middle of a glacier will pass over a much longer stretch than one at the margin during a given time. Consequently in a section through a glacier tongue, perpendicular to its movement, we shall find in its central zones much younger ice than on the sides.

But we will once more go back to the above mentioned stream lines which mark the path of a granule from its incorporation in the glacier to its final destruction and can be thought of as a channel of the diameter of the granule through which the latter and its successors have to travel.

In a glacier, the streamlines behave differently from those in a stream of water. When in a stream some obstacle, for instance a rock rising above the surface, hinders the flow, the water ascends on the upstream side and runs down on both sides of the obstacle. On the downstream side a concavity is produced into which the water flows from both sides until a short distance further down, the surface of the water again becomes level. Owing to the much lesser mobility of ice granules the process in a glacier requires much more time, or, expressed in terms of space, a much longer distance; in many cases even the level surface is not restored at all.

Now this is the point where the principal importance of our concept of horizontal age stratification within a glacier comes in. We know that the ice in the middle part of any vertical cross section of a glacier is much younger than in the lateral zones. Suppose than in a period of higher temperatures and reduced precipitation less snow accumulates in the névé basin than in the preceding years. Then evidently the outflow from the névé basin also will be reduced and in the central zone of the glacier tongue, where the younger ice flows, its thickness will diminish.

nish, while at the sides nourished by the ice of the previous fat years, the old level will be maintained (the effects of ablation being assumed to be uniform over the whole width of the glacier.)

Of course, ice will press in from the higher outside parts toward the center. But since near the glacier's surface the pressure is not strong, the transverse movement is slow and can not effectively counterbalance the lowering of the middle zone due to its undernourishment. On the other hand, surface streamlets of melt water will change their courses toward the deeper central parts. The reflection of the sun radiation in the concavity will raise the air temperature above the ice; water vapor which diminishes the energy of the radiation in both its directions can not be removed by the wind as completely as it was when the surface was plane or convex; and snow will more readily collect in the concave center.

These different processes clearly work in opposite senses and therefore compensate their effects fully or at least partially; consequently the impulse of the phenomenon, viz. the diminution of the alimentation out of the neve basin, doesn't become very much weakened or obscured.

The nature of the mentioned process has until now not been clearly recognized and duly described, as far as I know. E. Haug (Traité de Géologie, Paris, 1937, p. 418) writes :

« C'est à l'ablation, plus intense sur les bords que dans la région axiale, que doit être attribuée la forme convexe du glacier dans sa partie encaissée et à son extrémité inférieure. /Note : Sur plusieurs glaciers de l'Oisans, l'ablation est aujourd'hui si intense que, vers leur extrémité inférieure le profil en travers est non plus convexe, mais concave. / Dans la région du névé où l'alimentation prédomine sur l'ablation, les formes planes, voire concaves sont la règle. »

F. E. Matthes (Trans. Am. Geophys. Union, April 1946, p. 226) states, referring to the Elliot Glacier on Mount Hood, Oregon : « On the lower profile line, the surface was found to have been lowered by ablation during the past year by amounts ranging from one foot at the margins to five feet in the center. On the upper profile line, where the ice is only scantily covered with morainal debris, ablation had been more rapid and the surface appeared to have been lowered as much as 25 feet near the medial line of the glacier. A large share of this lowering, however, is very probably attributable to the erosive action of a streamlet of melt water. »

As long as the glacier is thick enough so that the sun radiation penetrating through the ice can not heat the ground, it can not be imagined why ablation might attack the middle parts more effectively, where the ice generally is thickest. When a central trough has been formed, the different secondary factors mentioned above begin operating and it is hardly possible to decide, whether they result either in deepening the depression or reducing its depth.

We will enumerate them once more; favorable to the lowering are : increase of the melt water inflow; increased effect of the sun radiation due to reflection; diminution of loss of heat due to minor removal of water vapor and hence lower effectiveness of outward radiation; — The opposite tendency have : inflow of the ice from the sides; greater deposition of snow in the hollow central parts; decreased effect of sun radiation due to water vapor.

If, on the other hand, a period of hot and dry weather is followed by one of low temperatures and increased precipitation, it is evident that the whole process must develop with contrary signs. The middle of the glacier tongue will swell and finally become convex; the younger ice here will press outward to the edges, to which the higher alimentation has not yet reached.

The reflection of the sun from the slopes of the valley has no more effect in producing convexity of the tongue during periods of advance, as it has in increasing the melting in the medial concave parts of the ice tongue during periods of wastage. *Both are phenomena of alimentation, not of ablation.*

The relation between the thickness of the glacier tongue and its extension down-valleyward is evident and has always been properly recognized. S. Finsterwalder told us, more than 20 years ago, when we determined the lowering of the surface of the Pasterze Glacier, of the Grossglockner group, in the Hohe Tauern, by means of levelling a transverse profile of the glacier tongue, that in consequence of it the glacier must become shorter, if the same conditions viz. the same amount of ablation, is maintained.

But it must be stressed that an increase of snow deposition must result at once in a considerable acceleration of the movement, much in the same way as high water in a river increases the velocity of its current.

H. Hess (Klimaschwankungen, Zeitsch. für Gletschk. 1941, p. 232) reports that in the Hintereisferner (Oetztaler Alpen, Tirol) the movement which in 1913-

1914 amounted to 30 meters per year, increased during the advance of 1914-1916 to 45 meters per year; during the advance 1917-1918 to 82 meters per year; and during the advance of 1918-1919 to 120 meters per year. But in 1921-1922, when recession set in, the velocity decreased quickly to 22 meters per year. On the other hand, the terminus did not advance until 1920 and it continued to advance until it reached a rate of 45 meters in 1923; then it began to recede.

It follows that there can be no doubt that the velocity of a glacier increases considerably during periods of higher alimentation, especially in the center of the glacier tongue. In consequence it becomes clear that *there is no better method of ascertaining, whether a glacier is about to increase or decrease in length, than by leveling a transverse profile of the névé line in different years.* Changes in the actual position of the terminus always must lag very much behind and moreover are affected by many factors in the local configuration of the glacier bed and local climatic circumstances.

The névé line itself is subject to change and moves to a higher position during dry periods. Hence leveling a transverse profile between two opposite fixed points anywhere below the névé line will do the same service; it will permit a prediction for as many more years to come, as it is located nearer to the névé line. Constrictions in the bed of the ice tongue must augment the amplitude of the surface oscillation, hence render it more evident. Even the simple observation of the center line of a glacier tongue in a longitudinal profile from the névé line downward and its comparison to the lateral parts of the surface may sometimes indicate, particularly in a regular glacier channel, whether a depression or swelling is propagating itself down from the upper parts. Since the flow of the ice through the glacier tongue takes a long time, different phases of the glacier's swelling or shrinking may occur at different altitudes along the length of the glacier at the same time. Determinations of the velocity of the glacier's flow and experiences will make it possible to predict the time of their arrival at the terminus.

Applying the above theory to the Derecho Glacier, in the Cordillera de Totora, of Argentina, we may state:

There is no doubt that the "rock island" exposed between the two arms of the upper part of the glacier, not long ago was covered by the ice. It became bare, when the middle part of the glacier was lowered during the dry period which has been observed throughout the entire Cordillera for many years back. The region below the rocks, overlain by the medial moraine which descends from the rock spur, is from 10 to 30 meters lower than the lateral surfaces.

The primary causes of the phenomenon are clear; but when the rock island rose out of the ice, secondary effects began to operate: The dark rock gets much more heated by the sun radiation than the clear ice; reflecting its heat during the day it contributes considerably to the melting of the ice. The two arms of the glacier, once separated, need more time and more space for their reunion, particularly because there is no pressure toward the glacier axis, since the flanks of the tongue have melted back from the slopes of the valley. The debris from the rock island, which formerly constituted a central moraine inside the ice, became a surface moraine. It is well known, that a thin layer of detritus accelerates melting while a thick one decreases or even stops it. It is therefore uncertain whether the new surface moraine influences in the sense of lowering the sunken central part of the glacier or contrarywise.

It is to be hoped that very soon a swelling and refilling of the medial trough will take place, which would announce an increase of the glacier's volume due to better alimentation during a climatic period of greater precipitation and cool weather, hence would be favorable not only to the moribund glaciers of the Cordilleras in the northern part of Argentina, but even in a higher degree to the water economy of the beautiful San Juan province.

San Juan, Argentina, September 10th, 1947.

STUDIES ON SNOW AND GLACIERS IN THE ARGENTINE REPUBLIC

by

EUG. ADOLFO A. MARCHETTI,

Chief of the Dep. of Hydrology of the National Meteorological Service
(Buenos Aires)

The object of the paper is to report to this Congress on the work recently performed by the Department of Hydrology of the Argentine Meteorological Service concerning studies on snow and glaciers of the patagonic region, i.e., the Southern part of the country, including also the provinces of Mendoza and San Juan. It is considered that those two elements — snow and glaciers — are the feeding sources of the numerous rivers which run through the region having the basin in the mountains (Cordillera de los Andes) and most of them flowing into the Atlantic Ocean. In order to perform such a study it has already been accomplished numerous permanent installations, namely :

34 Rainfall and snowfall gauges.

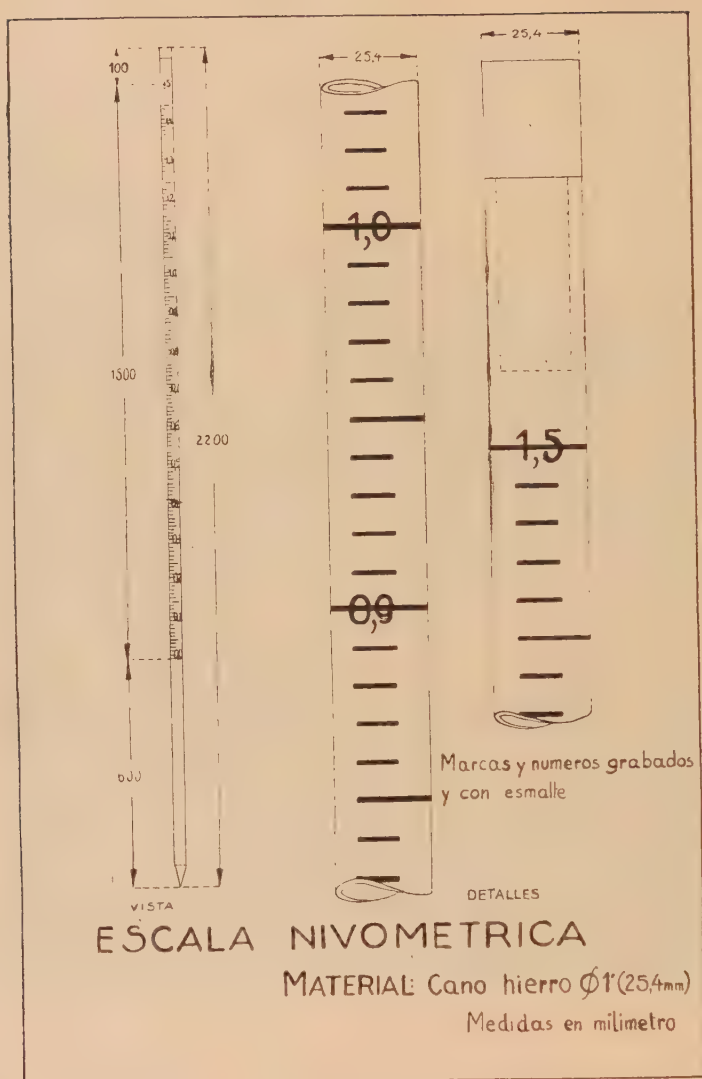
20 Daily reading snowmeters, where it was possible to find observers, since we are dealing with almost depopulated regions.

Simultaneously it has been faced the hydrometric study of the rivers feeded by snow and glaciers and established 58 permanent stations to gauge them. We have already performed numerous measurements of volume and evaluated the corresponding curves as in the case of Santa Cruz river (one of the most important ones of that territory), Rio Colvado, Rio Grande, etc.

The hydrologic study of the patagonic region is a problem of fundamental importance to the Argentine Republic and its solution has to be attained as soon as possible due to the fact that the fulfilment of a vast plan of hydraulical works and constructions of roads and railroads which are going to contribute to the progress of such an extended and rich zone rely upon it.

The paper is divided into two chapters showing in detail the plan of studies which has been followed so far.





REGISTRO DE OBSERVACIONES PLUVIOMETRICAS

19

Lugar

Prov. o Terr.

1		2		3		4		5		6		RESERVADO PARA LA OFICINA CENTRAL			
DÍA	Hora de la observación	PRECIPITACION		Cantidad de lluvia o nieve caída en 24 horas en mm.	Altura de la nieve sobre el suelo en cm.	Lluvia	Nieve	Granizo	García	Nebulina	Tormenta de truenos	Hielada	OBSERVACIONES (Dirección e intensi- dad del viento y otras indicaciones)		
		Empeto a horas	Terminó a horas										T	D	I
1															
2															
3															
4															
5															
6															
7															
8															
9															
10															
11															
12															
13															
14															
15															
16															
17															
18															
19															
20															
21															
22															
23															
24															
25															
26															
27															
28															
29															
30															
31															
SUMAS															
Medición de las hs. del día 1.º del mes siguiente mm.															
Llovió desde horas del día hasta hs. del día															

Setlo de la oficina o estación a cargo
del aparato y observaciones.

Profundidad de la napa el último día
del mes, en metros y centímetros

Certifico que los datos de esta planilla son exactos.

FECHA

FIRMA

Form. H-760

MINISTERIO DE AGRICULTURA
DIRECCION DE METEOROLOGIA
GEOFISICA E HIDROLOGIA
DIVISION HIDROLOGIA

PRECIPITACION MEDIDA CON TOTALIZADORES

.....
 Lugar..... Prov. ó Terr..... Estación.....

Latitud..... Longitud..... Altura sobre el nivel del mar.....

[illegible]

Pesos: Recipiente: R..... Aro superior: S..... Aro inferior: I.....





8) DER WASSERABFLUSS IN EINEM GLETSCHERTAL

von

Dr. R. STRIFF-BECKER Zurich.

Um den Einfluss eines Gletschers auf den Wasserhaushalt eines Tales festzustellen, müssen drei Abschnitte der Vergletscherung betrachtet werden :

- I. Die Gipfelzone von den Gipfeln und Graten abwärts bis zur Höhe der Bergschrunde
- II. Die Zone zwischen Bergschrund und Firnlinie.
- III. Die Zone von der Firnlinie bis zum Ende der Gletscherzunge.

I. Die Gipfelzone

Hier fällt der Niederschlag auch im Sommer fast ausschliesslich in fester Form als Schnee. Der selten vorkommende Regen versickert im lockeren Hochschnee und gelangt kaum zum Abfluss. Etwas fliessendes Wasser entsteht durch Schneeschmelzung auf dunkelfarbigem Felsen bei Sonnenschein. Auch dieses Wasserquantum ist gering und gefriert zudem nachts grösstenteils wieder, ehe es den Grund des Gletschers erreicht. Am unteren Rand der hochgelegenen dicken Firnpolster der Hänggletscher sieht man kein Wasser ausfliessen, obgleich dort im Inneren des Eises bereits kriechende Bewegung vorhanden ist und etwas Schmelzung infolge des Massendruckes und der Bewegung in Betracht kommt. Frisch abgestürzte Firneistrümmer sind trocken und die höchsten Firnspaltenwände weisen trockene Flächen auf. Diese oberste Zone I kommt demnach für direkte Speisung des Gletscherbaches in flüssiger Phase kaum in Betracht.

II. Die mittlere Zone bis zur Firngrenze

In dieser zum Nährgebiet des Gletschers gehörenden Zone zwischen Bergschrund und Firnlinie fällt neben Schnee, Graupeln und Hagel, schon häufiger auch Regen. Dazu kommt Schmelzwasser infolge höherer Lufttemperatur und stärkerer Luftfeuchtigkeit. Beide Ursachen der Wasserbildung nehmen talwärts progressiv zu. Ein Teil des Regen- und Schmelzwassers sickert im Firneis bis zu variabler Tiefe ein, bleibt da kapillar festgehalten und gefriert in den oberen Schichten nachts zu Eislamellen, zu Harteisschichten verschiedener Dicke und Ausdehnung. Diese wiederum gestatten den Abfluss neuen Schmelz- oder Regenwassers längs der Oberfläche oder unterirdisch. Man sieht im Frühsommer häufig an der Firnoberfläche parallele Furchen in den Gefällsrichtungen. Das deutet darauf hin, dass unter jeder Furche ein Wasserlein auf der Hartschicht in der Gefällsrichtung abfließt, mit anderen sich vereinigt bis das Wasser Gelegenheit findet, durch Spalten und Strudellöcher auf den Grund des Eises zu gelangen und den Anfang des unterirdischen Gletscherbaches zu bilden. Man hört oft in frühen Morgenstunden, wenn an der Oberfläche der Firn infolge nächtlicher Ausstrahlung hart gefroren ist, im Inneren des Eiskörpers das Gurgeln fliessenden Wassers vom warmen Vorabend her. In dieser zweiten Zone beginnt auch die Abschmelzung an der Sohle der Firnmasse.

Im Sommer steigt die Nullgrad Isotherme des Felsbodens langsam bergwärts. In der Nähe dieser Temperaturgrenze genügt nach unserer Erfahrung eine Firnschicht von 25 bis 30 Metern Dicke, um die Firnkörner des bodennahen Schichten fluid-plastisch zu machen unter dem Einfluss der konstant wirkenden Erdwärme, sodann auch durch das Gewicht der hangenden Firnmasse. Der Betrag der jährlichen Abschmelzung ist zwar, wie berechnet werden kann, nur gering. Im Inneren der tiefen Firnschichten muss auch etwas Wasser durch die Reibung infolge der Bewegung gebildet werden. Nach Ansicht des Autors (*) fließt das Firneis infolge des Druckes der überlagernden Massen in den tieferliegenden Schichten schneller, als nahe der Oberfläche.

(*) R. Streiff-Becker, « Zur Dynamik des Firneises », Zeitschr. f. Gletscherkunde Bd. XXVI, 1938.

Diese Fliessbewegung muss eine differenziale sein, ähnlich dem Fliessen körniger Massen in einem technischen Silo. An den Oberflächen der einzelnen Firnkörner tritt an gewissen Berührungsstellen mit benachbarten Firnkörnern schleifende Reibung auf und durch Reibungswärme etwas Schmelzwasser, das als Schmiermittel wirkt und die Bewegung erleichtert. An gewissen anderen Stellen derselben Körner kommt gleichsinnige, rollende Reibung vor, d.h. Druckentlastung und Regulation. Beim Bewegungsvorgang wird so etwas Wasser erzeugt, jedoch daneben wieder in Eis zurückverwandelt, zumal in den höheren, kälteren Lagen. In geringerer Meereshöhe, näher der Firnlinie, mag bei der Firnbewegung auch überschüssiges Schmelzwasser erzeugt werden, sodann in zunehmender Menge Schmelz- und Regenwasser von der Oberfläche her, und drittens etwas Schmelzwasser von der Sohle durch Erdwärme. Der Gletscherbach nimmt daher in dieser II. Zone seinen Ursprung, kommt jedoch erst in der nun folgenden III. Zone im eigentlichen Ablationsgebiet des Gletschers, zur vollen Bedeutung.

III. Untere Zone von der Firnlinie bis zum Gletscherende

Diese Zone ist Hauptlieferant von Wasser für den Gletscherbach. Die Abschmelzung geht im Sommer, entsprechend der Witterung, von der Sohle bis zur Oberfläche, Tag und Nacht, mehr oder weniger kräftig vor sich. Bei Regenwetter fliessen von den beidseitigen Talhängen die Wasser in zahlreichen Rinnsalen gegen den Gletscher, stürzen in die Randklüfte, und gelangen unter dem Eis durch zum glazialen Bach. In den meisten Fällen verhindern jedoch die Seitenmoränen den direkten Zutritt zum Gletscher, so dass dieses Hangwasser erst ausserhalb der Seiten- u. Stirn-moränen den Gletscherbach erreicht.

Im Winter ruht in allen Zonen die Oberflächenschmelzung fast gänzlich, ebenso die Schmelzung durch die Erdwärme an den Gletscherrändern soweit als die Dicke der Eisaufgabe nicht genügt, um das Eindringen der kalten Aussentemperaturen zu verhindern. Unter dem dickeren Eispanser der höheren II. Zone wird auch im Winter ein kleiner Wasserabfluss möglich sein, erstens durch Schmelzung infolge der konstant wirksamen Erdwärme, sodann ein wenig durch die Bewegungswärme, weil das Firneis im Winter schneller fliesst, als im Sommer.

In einem Gletschertal können auch subglazial vorkommende Quellen als Wasserspender in Betracht kommen, doch kann das nur der Geologe entscheiden.

Gletscher und Firne sind zwiespältige Regulatoren des Wasserabflusses, d. h. sie können unter Umständen günstig wirken, unter anderen Umständen sehr ungünstig: z.B. günstig durch Retention des fallenden Niederschlages im lockeren Firnschnee, oder beim Uebergang von Regen- zu Schneefall. Auf dem relativ warmen, aperen Boden schmelzen anfangs die nassen Schneeflocken und speisen den Bach, während sie auf dem kalten Firneis sofort als künftige Wasserreserve liegen bleiben. Umgekehrt kann zu anderen Zeiten der Gletscher, anstatt retardierend, auf den Wasserhaushalt des Baches ungünstig einwirken. Das ist besonders der Fall an heissen, gewitterigen Tagen. Da kommt zum grossen meteorischen Niederschlag gleichzeitig das viele schon vorhandene Schmelzwasser hinzu. Vermögen die engen Querschnitte der Wasserläufe im Inneren des Gletscherkörpers die Wassermengen zu schlucken, wird der Abfluss ohne Schaden stattfinden, im anderen Fall kann eine Katastrophe eintreten. Es kann vorkommen, dass die Kanäle die Wassermenge nicht zu schlucken vermögen oder gar durch mitgeführten Schutt verstopft werden. Da kann es durch Rückstau im Inneren des Eises zu hohem, viele Atmosphären betragenden Drucke kommen und zu plötzlichen Durchbrüchen, ohne dass man gleich an einen See im Innern des Gletschers denken muss. Grössere Hohlräume sind wohl möglich, z.B. da, wo ein Schmelzbach von der Oberfläche in einen Eisschacht zum Gletscherinneren stürzt und dort durch Rückprall Höhlungen ausstrudelt.

Die Erosionskraft des Wassers ist in der Gipfelzone I nahezu gleich Null, nimmt in der II. Zone progressiv zu und erreicht in der III. Zone ihr Maximum, unterscheidet sich jedoch in der Wirkung von derjenigen des Wassers in einem gletscherlosen Tal. In Letzteren fliesst während längeren Trockenperioden kein Wasser mehr, ebenso keines im Winter bei strenger Kälte, sofern man vorkommendes Quellwasser nicht berücksichtigt.

Im Gletschertal fliesst auch im strengen Winter stets etwas Wasser herrührend von der Abschmelzung an der Sohle des Eises dicker Gletscher durch die Erdwärme, im Sommer sowieso durch Eisschmelzung an der Oberfläche. Die Wasserführung

ist zur warmen Jahreszeit sehr verschieden stark, gleichmässiger bei warmer, trüber Witterung, wenn die Abschmelzung auch bei Nacht andauert, intermittierend bei hellem Wetter, weil infolge der nächtlichen Ausstrahlung das Wasser an der Oberfläche grösstenteils gefriert. Der Gletscherbach ist dann in den frühen Morgenstunden klein, schwillt gegen Nachmittag, wenn die Abschmelzung ihr Maximum erreicht, rasch an, und kann, wenn ein allfälliges Wärmegewitter hinzutritt, gefährliche Dimensionen annehmen.

Im gletscherlosen Tal wird der mitgeführte Schutt nach unten hin bald weniger erosionsfähig, weil die einzelnen Komponenten sukzessive abgerundet werden. Im Gletschertal führt das Eis auf seinem Rücken scharfkantige Gesteine talwärts mit sich. Es kann so in jeder Höhenlage scharfkantiges Material in den Gletscherbach gelangen, welcher zuweilen infolge hohen Druckes, wie oben erklärt, mit grosser Geschwindigkeit durch die engen Kanäle strömt, eine kräftige subglaziale Erosion bewirkt und tiefe Löcher und Rinnen auskolkt, während gleichzeitig das langsam kriechende, schleifende Eis die Zwischenräume zu Rundhöckern umformt. Eine wichtige Rolle spielen dabei nicht nur die Härtegrade der mitgeführten Gesteine sondern auch das Quantum der Steine und Sande. Ein Uebermass von Schutt führt leicht zu Verstopfungen, zu Ruckstau und bei steigendem Druck zu gewaltsamen Durchbrüchen und Aenderungen der subglazialen Abflussrinnen.

In Zeiten starker Gletscherrückzüge kann es vorkommen, dass ein Stück Gletscher als sog. Toteis vom Hauptgletscher abgetrennt wird. Dieses Toteis kann bei genügender Schuttbedeckung jahrzehnte lang erhalten bleiben und ebenso lang ein Abschmelzbächlein, einer Quelle gleich, nutzbringend liefern.

Fast auf der ganzen Erde macht sich eine starke Schrumpfung der Firne und Gletscher, besonders deutlich seit 1927, bemerkbar, durch Rückzug der Zungenenden, Abnahme der Eisdicke und Abnahme der Firnnährflächen, wodurch ein grosser Verlust der nutzbaren Wasserreserve entsteht, welche zum Ausgleich der Wasserführung in Trockenperioden so wichtig ist.

Wie gross des Massenverlust an Firn- und Gletschereis durch Verdunstung, Sublimation, ist, kann mit Sicherheit kaum gesagt werden, weil in dieser Hinsicht noch viel zu wenig systematische Untersuchungen gemacht sind.

Zürich, März, 1948.

SUMMARY

In a glaciated valley we can discern three zones supplying the glacier stream with water.

I. The upper zone from the peaks to the bergschrund, which furnishes almost no water. Rain is very seldom and is absorbed by the deep dry snow. The little water produced by melting during the day freezes again at night.

II. The middle zone from bergschrund to the firn line is the birthplace of the glacier stream. Rain and water from melted snow and ice are sucked down into the firn snow, forming layers of black ice beneath the surface, on which the water melted during the day runs downwards, sometimes falling into crevasses or glacier mills, and forms the glacier stream on the glacier bed. In addition, a little water is produced from ice melting in consequence of the warmth of the earth and the movement of the ice; this occurs as soon as the ice is more than 30 metres thick (in the Alps).

III. The lower zone from the firn line to the end of the glacier tongue furnishes water throughout the whole year.

Even in winter water runs from big glaciers because of heat from the earth and the ice movement in the middle zone II, where ice movement is rather in winter than in summer, will produce some water. Rain water from the surrounding slopes runs through marginal crevasses to the subglacier stream, mostly however beyond the terminal moraines. In summer water is abundant. To the rain water is added the water melting from the surface. At night the amount of water formed by melting is much less, while during the day it increases. Should a thunder storm with much rain occur on a very hot day when much water has been produced from the melting ice, it is possible that the channels within the glacier cannot absorb all the water, so that hydraulic pressure increases and the ice mantle may burst and thus cause a devastating flood. The erosive power of a glacier stream is considerable since the water from a glacier is loaded with sharp detritus.

9) THE PRESSURE MELTING POINT OF ICE, AND THE EXCAVATION OF CIRQUES AND VALLEY STEPS BY GLACIERS

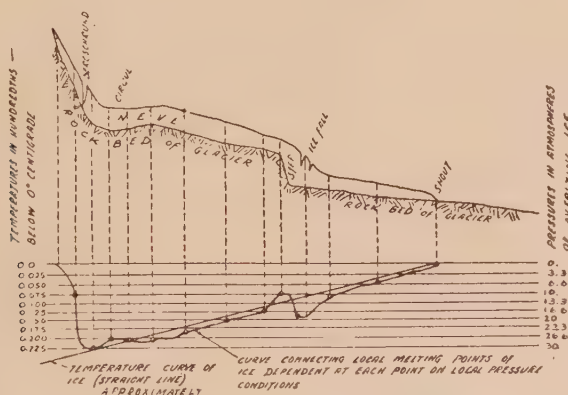
by

Joel E. FISHER
25, West 43rd Street, New-York 28, N. Y.

Recent studies (Monthly Notices, Royal Astronomical Society, Geophysical Supplement) Vol. 4. 1939, p. 617, T. P. Hughes and G. Seligman) have developed a now well established principle, that outside the arctic the temperature of all glaciers 10 or 20 meters below the surface (and deeper) tends to be at all times at the local pressure melting point; that is, the temperature is continually very slightly below 0° Centigrade (Journal of Glaciology, Vol. 1, 2, 1947, p. 61, article by W. V. Lewis); (see also Seligman, G., The Structure of a Temperate Glacier, Geographical Journal, Vol. 47, 1941, p. 295-317). The upper 10 or 20 meters of ice, affected by colder winter temperatures, may cool down somewhat lower, seasonally, each winter).

This principle requires that the temperature at the very bed of any glacier, as one follows that bed, up-glacier, shall gradually drop from exactly 0° Centigrade (at the snout) to temperatures fractionally lower, up-glacier, as the pressure of overlying ice and névé irregularly increase. This « pressure » is, primarily, that of the static load of vertically overlying ice and névé, but it may, and often does include some measure of the hydraulic head of the glacier upstream; because, under the extrusion flow process of glaciers, the brittle surface ice, joining in with the rock walls and bed, provide a sort of pipe, within which the flowing ice behaves to a degree like a liquid, and therefore exerts a pressure (measured by its up-glacier head), in all directions, including vertically down wards.

The temperature of the bed ice, under 500 meters of overlying névé, alone with a static pressure of some 700 lbs. per square inch, must evidently be lower than, the temperature of any ice near the snout, else, if temperature of the ice everywhere on its bed were exactly 0° Centigrade, melting would be greater under the heavier pressure of thicker beds of névé, up-glacier than at the snout. Actually, the melting point of ice is lowered .0075 $^{\circ}$ Centigrade per additional atmosphere of pressure.



Successively colder temperatures below the surface up-glacier are altogether logical, as (except for localized precipitation), all glacier ice and névé derive originally from the snows falling above the bergschrunds, and those upper snows and ice originally possessed a temperature well below 0° , the year round. (Monthly Notices, Royal Astronomical Society, Geophysical Supplement, Vol. 4, 1939, p. 631, T. P. Hughes and G. Seligman). Witness, the aprons of ice, frozen solid to the

surrounding walls of rock, above the bergschrunds on any hardy Alpine peak. Noon-time melt water observed on such upper snow fields *above bergschrunds* is only a superficial surface feature and should not be viewed as indicative of temperature, below the immediate surface, warm enough to permit melting, else, the bergschrunds would close up! In fact, it is possible to say that bergschrunds mark the boundary between snow or ice whose temperature is *below* the pressure melting point (above the bergschrunds) and snow or ice whose temperature (1) is *at* the local pressure melting point (below the bergschrund). This tie-in of any bergschrund with climatic altitudes explains the much discussed « Cirque-Niveau » in any given district.

From these propositions, it is reasonable that the temperature of the ice (névé) on the bed of any glacier, *up close to the bergschrund*, must be just *at the local pressure melting point*, (2) but because that very same ice (névé) was only very recently part and parcel of the upper regions of snows (above the bergschrund) which are definitely *below* their local pressure melting point, such névé has no past history of existence *at the pressure melting point*; it cannot yet have accumulated, within its capillary passages, any *substantial* amount of melt water — no more than *just* enough to assure that its contact with the bed rock will be a moderately wet, lubricated contact as opposed to any frozen « adherence » to the rock, as is found above bergschrunds. Thus, any rocks held within that recently born névé ice will be held in a matrix which carries so little water within its voids as to minimize any retreat under pressure, within its own icy medium, of those tools which it carries, even though they may be held against the bed rock under the heavy pressure of the weight of thick overlying névé. (Attention is directed to the particularly greater thickness of overlying snow at the uppermost limits of any névé field due to aprons of snow reaching steeply up the cirque slopes to the bergschrunds).

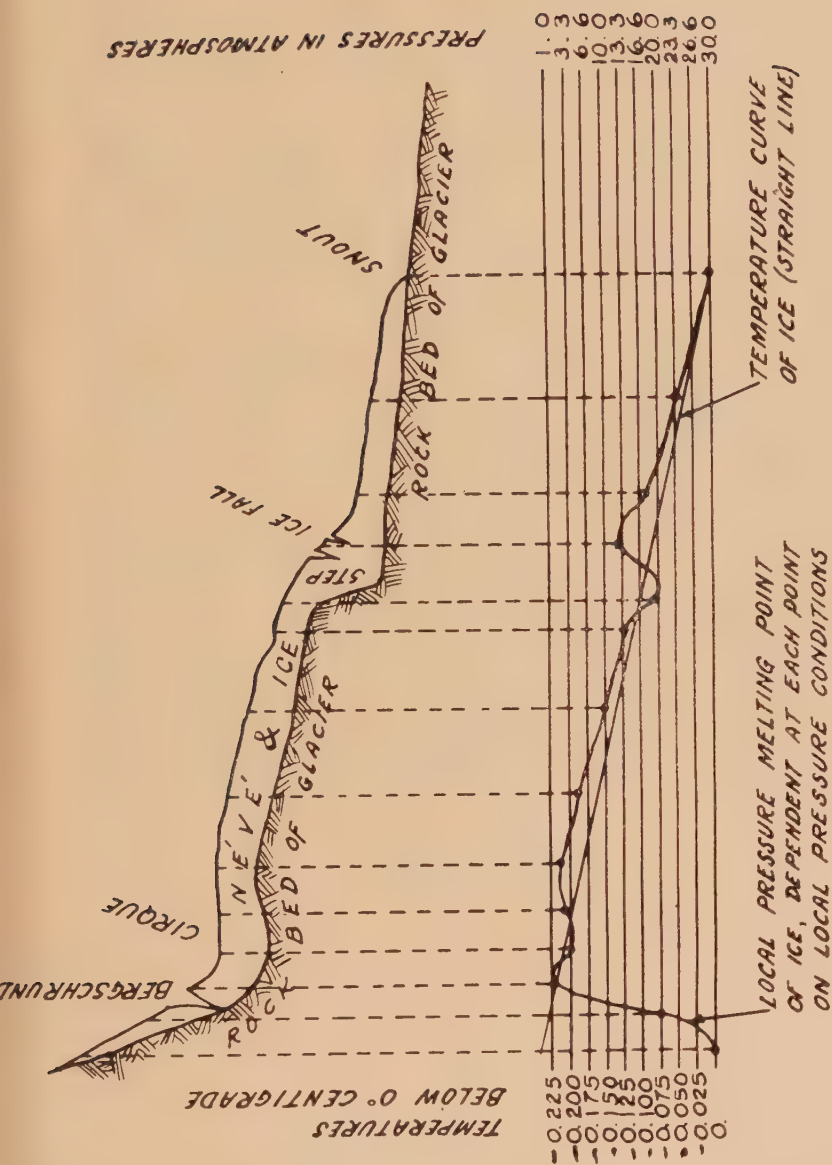
Thus the temperature of the basal layer of ice, up close to the bergschrund, regardless of however heavy a static load it may carry, is *just* over the line of being above the local pressure melting point: *just* able to provide a wet, lubricated bed; *just* able to cause the ice to yield, minutely, to pressure; not, however, sufficiently over the line of the pressure melting point to permit any rocks held in that ice to be pushed back readily into its icy medium. It will be recognized that the latent heat required for any minute local « melting » will always delay any quick transition from « hard » to « soft » ice, as a result of quickly increased pressure, so that for some time any ice, although under increasing pressure, is still very hard rigid ice. There is thus a triple combination which exists only *immediately below the bergschrund* — (a) presence of vigorous cutting tools freshly fallen from the heights through the open bergschrund, to the bed of the glacier, (b) extreme static pressure thereon, which generates, right up at the foot of the head wall of a cirque, a particularly vigorous cutting action, (c) ice temperatures *barely* above to the pressure melting point and therefore holding such tools in an extremely firm grip against the bed. It is this triple combination which is responsible for the excavation of cirques, and of these three factors, it is the third, the relationship of ice temperature to local pressure melting point, which is a feature unique to the physics of glaciers. Against these three factors, velocity of motion is insignificant, provided only that there is *some* motion. We are too prone to impute to faster motion of ice a correspondingly faster supply of cutting tools. That is not so. The number of cutting edges carried across the bed by a glacier depends altogether on the rate of supply of cutting tools, not on the velocity of the glacier itself.

Consider next what changes occur in the characteristics of the ice as it ages and moves down stream from the cirque wall.

Except for a thin veneer of surface ice, the glacier can never become any colder, for winter temperatures do not penetrate below 20 meters and mean summer temperatures are well above freezing; on the other hand, the ice can pick up some heat by conduction from melt water seeping down, and, very slowly, from its bed rocks which conduct heat from within the earth; and by friction, too. If heat from these sources is insufficient to warm the ice enough to keep pace with a rising pressure melting point (rising, due to gradual reduction in pressures as the ice thins out down-valley), flowage slows down, erosion of the bed increases, the névé reaccu-

(1) This statement refers, of course, to the main body of névé, not to the superficial upper crust where temperatures during winter may fall below the pressure melting point.

(2) In the oft-reported classic exploration of a bergschrund on Mt. Lyell by the late W. D. Johnson, in which he told of water permeating its up-hill face, that water is to be thought of as merely surface melt water flooding down into the bergschrund from above, and not for a moment, as water exuded from capillary passages within the bed of the suprabergschrund ice slope, from regions 20 meters or more below its surface, regions therefore unaffected by surface air temperatures.



CURVES OF DECREASING ICE TEMPERATURES AND OF LOCAL PRESSURE MELTING POINTS, SUPERIMPOSED. WHEREVER FORMER LIES ABOVE LATTER ICE IS COLDER THAN THE LOCAL PRESSURE MELTING POINT. IT IS HARD ICE, FLOWAGE IS RETARDED, EROSION OF BED IS INTENSIFIED. WHEREVER FORMER IS BELOW LATTER, MELTING STARTS, ICE BECOMES SOFT, FLOWS FREELY, EROSION AT BED IS LESS EFFECTIVE

mulates deeper and deeper until its thickness is again sufficient to build up a sufficient static pressure to place it once again on the warm side of the curve of pressure melting points. Once this equilibrium point is regained, the ice in the glacier's bed will begin to accumulate ever so little more water in its capillaries; to accumulate heat sufficient to place it a little further over the equilibrium of local pressure melting points; once this begins to happen, while the ice is still a far cry from any condition of spongy, cheesy ice, it no longer has *quite* the viselike characteristics it formerly had; the cutting tools in its bed no longer are *quite* so insistent on gouging out the bed rock and, also, those glacier worn cutting tools are no longer *so* sharp. Here, then, in this slight relaxation of pristine erosional vigor, we can see just how the downstream lips of corrie tarns first induced into existence. The reader at this point is referred to the interesting studies of Hans Carol (*Journal of Glaciology*, Vol. 1, #2, 1947), on Roches Moutonnées.

This very same process may repeat itself any place farther down the glacier's course, wherever initiated by a steep drop in the grade of the bed (due, perhaps, originally to structure). Against the back wall of such an incipient step, the same three factors thus exist which accounted for cirque excavation — or for tarn excavation — unworn cutting tools (derived now from the back wall of the step itself); locally increased static pressure of ice on its bed; and (to a lesser degree, but still true) temperatures of ice colder locally than local pressure melting point.

It will be noted that a rise in the pressure melting point occurs first at the top of the step. This ice, now hardened, as it goes over the step, is soon subjected to greater static pressure, but this greater pressure cannot instantly resoften the ice because the latent heat absorbed in minute local melting operations creates a lag. Pressure can thus rejuvenate cutting power, appearing as excavation at the foot of the step, just as at the headwall of a cirque; eventually, this greater pressure overcomes this lag in softening the physical state of the ice — the ice softens again, and the extreme vigor of its erosion ability ends once more.

In short, while *temperature* of the ice near the bed of a glacier follows a near straight-line graph, rising slowly as one down glacier, *pressure* at selected points may vary greatly, depending on special local conditions. On which side of the curve of equilibrium of pressure melting points the temperature of any particular portion of the ice may find itself, will therefore vary, often quite quickly, with resulting changes in the « state » of the ice and in the cutting power of tools embedded therein, as the glacier moves through regions of varying pressures.

The processes of cirque excavation, of tarn excavation, and of step cutting by glaciers, with their apparent anomalies of reverse slopes persisting in profiles, thus are seen to stem from the irregular curve of the local pressure melting point of the ice, locally superimposed on a nearly straight line gradual rise in the actual temperature of the bed ice, the latter starting from a temperature definitely below any possible pressure melting point (above the bergschrund) and rising all the way to a point (at the snout) which is exactly *the* pressure melting point of ice at atmospheric pressure.

10) PRESSURE MELTING POINTS OF ICE AND THEIR CONTROL ON THE PROFILE OF GLACIATED VALLEYS

by

Joel E. FISHER

25, West 43rd Street, New York 28, N. Y.

Assuming that the temperature of ice, 20 meters or more below the surface of all temperate zone glaciers is always at the pressure melting point of ice at the point under consideration, and assuming that the vertical thickness of ice is, in general, greatest, up near the lower lip of the bergschrund, at the head of the cirque, the conclusion is reached that the actual temperature of ice, close to the bed of a glacier, is minutely lower than 0°C . in the upper reaches of a glacier, because the temperature of the melting point of ice decreases with increasing pressure.

Inasmuch as the temperature of the ice is exactly 0°C at the snout, it follows that the temperature of ice in a glacier must decrease, gradually, from its snout, going up-glacier. Within reasonable limits the plot of this temperature may be drawn as a straight line, on cross-section paper where the Y axis measures negative temperature, and the X axis measures stations along the path of the glacier.

Another graph, representing the static pressure at the bed of the glacier at each station, may be superimposed on this plot; that pressure graph will further reflect hydrostatic head of the glacier, where extrusion flow processes obtain; and that pressure graph will obviously be a sinous curve; if plotted to suitable scale, it can be so drawn as to cross and recross the temperature plot; each crossing marking a point where actual temperature of ice coincides with local pressure melting points.

Every area where the pressure curve is below the temperature plot will thus mark a region where the physical state of the ice is minutely over on the cold side of equilibrium of water and ice, where there is no melting, but regelation; conversely, every area where the pressure curve is above the temperature plot will mark a region where the physical state of ice is minutely on the warm side of equilibrium, where melting will develop.

Bearing in mind that the substantial latent heat involved in transition from water to ice acts as a heavy brake on any mass transition, retarding it by the slow conductivity of heat through ice to or from the outside, it nevertheless follows that the bed of the glacier will consist of ice, in some areas, on the soft side, in others on the hard side of ice at the exact pressure melting point equilibrium; cutting tools where overlaid by the first state, will have little cutting power, regardless of the static load overhead; cutting tools in the other hard area will have active cutting power.

The changes in this cutting power are thus correlated with the characteristic features of the profile of a glaciated valley-cirques, steps, etc. and are shown to be supported by this relationship of pressure melting point to any originally slight variations in depth of ice overlying such areas.

The existence of the so-called cirque-niveau, in any one region, is also explained on the same principle, integrated with locally constant climatic altitudes.

MARCH 24, 1948.

11) LES SONDAGES SISMIQUES DE LA COMMISSION HELVETIQUE DES GLACIERS S.H.S.N.

RAPPORT A L'ASSOCIATION INTERNATIONALE D'HYDROLOGIE

établi par

A. RENAUD et P. L. MERCANTON

Membres de la Commission helvétique, pour la session d'Oslo, de l'U. G. G. I., août 1948.

I. — Préambule

La connaissance de l'épaisseur des glaciers et de la configuration de leur lit a une importance très grande pour la science et l'économie. La glaciologie, la géophysique et l'art de l'ingénieur sont directement intéressés à ce problème. C'est une des grandes préoccupations de la Commission helvétique des glaciers (C. H. Gl). Avant d'exposer la contribution que celle-ci a apportée à la solution du problème, nous rappellerons brièvement les travaux antérieurs et leurs principaux résultats.

II. — Les sondages mécaniques

La première tentative de sondage mécanique est due à l'initiative du Suisse *Louis Agassiz* qui réussit en 1842 à forer un trou profond de 65 m dans le glacier inférieur de l'Aar, sans toutefois atteindre le fond. Il fallut attendre jusqu'en 1895 pour que la méthode des forages fut reprise et améliorée par MM. *Hess* et *Blümcke* qui l'appliquèrent dès lors avec un succès complet à l'Hinterseisferner (Autriche). Entre 1895 et 1909, ces deux savants forèrent 11 trous qui atteignirent le lit du glacier à des profondeurs comprises entre 66,5 m et 224 m. Ces résultats remarquables firent date dans l'avancement des études glaciaires: les trous forés furent utilisés pour des observations sur la température interne du glacier et pour diverses observations qui contribuèrent à confirmer expérimentalement la théorie cinématique de l'écoulement, de *Finsterwalder*. A la même époque, et d'une façon indépendante, MM. *C. Dutoit* et *P. L. Mercanton* firent un essai de forage au glacier du Trient (Suisse) par le même procédé. A la même époque encore, le Service forestier français fit sonder de la même façon et avec un plein succès, le glacier de Tête Rousse (Mont Blanc).

Toutefois, en raison de son coût élevé à l'époque, le forage mécanique n'eut guère de continuateurs et ses applications restèrent limitées au percement de trous de quelques mètres de profondeur, destinés à loger des perches pour des mesures d'ablation. Il a été remis en honneur récemment par le Service de l'Electricité de France, par *M. Ract-Madou*, ingénieur, qui l'appliqua avec succès à la prospection de la langue du glacier des Bois (massif du Mont-Blanc), sans toutefois dépasser une centaine de mètres.

D'autre part, et au début de ce siècle déjà, on avait essayé la perforation thermique du glacier par un corps réchauffé par le courant électrique, sans succès. Durant la dernière guerre en revanche, *M. Calciati* (Italie), en utilisant une tête de sonde chauffée par un courant d'eau bouillante, a réussi une série de forages jusqu'au lit du glacier de Hohnsand (119 m), poussant le sondage à raison de 3 à 4 m. par heure en moyenne, exceptionnellement jusqu'à 10 m. par heure.

III. — Les essais de prospection physique

Les forages mécaniques sont longs, aléatoires et dispendieux; de tous côtés donc et tout spécialement en Suisse, on chercha un procédé de prospection glaciaire plus pratique, une méthode d'ordre physique. *Mr Mercanton* fit en 1926 des essais avec des explosifs et des géophones, puis tenta déjà, d'autre part, d'utiliser les ultrasons. Ces tentatives furent toutes infructueuses, mais il convenait de les rappeler pour montrer l'importance que la C. H. Gl. accordait à ce problème et sa volonté d'y trouver une solution.

En 1927, il appartint à M. *Mothes*, de l'Institut de Géophysique de Göttingen (Allemagne), de procéder à des essais qui inaugurèrent une nouvelle ère mémorable de la glaciologie en appliquant avec succès les méthodes de prospection sismique au glacier d'Hinterseifner (Autriche). A l'aide d'un sismographe très sensible, il réussit en effet pour la première fois à enregistrer utilement les ondes qu'une explosion provoque dans la glace et à en déduire l'épaisseur du glacier. La parfaite concordance de ses résultats avec ceux des forages de MM. *Hess* et *Blümcke* attira l'attention des glaciologues qui s'empressèrent dès lors de tirer parti de cette méthode. Après de nouveaux essais au glacier de Pasterze (Autriche) et au Grand Glacier d'Aletsch (Suisse) en 1929, les géophysiciens allemands furent en mesure de fournir à A. *Wegener* un matériel sismométrique qui fit ses preuves sur l'Inlandsis Groenlandais (1929-30) dont la détermination de l'épaisseur (on trouva de 2200 à 2700 m à Eismitte), devait éclairer un des chapitres les plus intéressants de la géophysique. De son côté, la C. H. Gl. procéda en 1931 à des essais encourageants sur le glacier du Rhône (Suisse), en collaboration avec les géophysiciens de Göttingen. Ce succès décida la C. H. Gl. à rechercher en Suisse le matériel et les spécialistes d'une méthode qu'elle se proposait d'appliquer systématiquement à nos glaciers.

IV. — Le Sismographe de la Commission Helvétique des Glaciers

Les intentions de la C. H. Gl. furent réalisées avec un rare bonheur par M. A. *Kreis*, professeur de physique à Coire (Suisse), qui construisit pour elle un sismographe remarquablement adapté à la prospection du glacier. Il s'agit d'un instrument portable, d'un poids très réduit, la masse stationnaire ne pesant que 12 kg. Il donne les trois composantes du mouvement, verticale et horizontales, avec une amplification mécanique et optique de 30 000 fois environ. La mesure du temps y est obtenue par l'emploi d'un diapason électro-magnétique à 50 périodes par seconde commandant l'illumination concomitante d'un tube contenant de l'hélium. L'image du trait de lumière ainsi déclenché par chaque vibration du diapason est enregistrée sur une bande de papier au bromure d'argent dont la vitesse de déroulement permet d'apprécier le temps à 0,001 s. près, au moins. La même bande (fig. 1) enregistre l'impulsion donnée à l'équipage mobile d'un petit oscillographe à l'instant de l'explosion, ainsi que les diverses ondes élastiques produites par celle-ci : ondes longitudinales directes P et réfléchies R, ondes transversales directes S, ainsi que d'autres, par exemple celles qui, réfractées au contact du lit rocheux et transmises par ce dernier, présentent dans certains cas le plus grand intérêt pour cette prospection. Les ondes P ont un spectre dont on peut estimer la fréquence comme comprise entre 5 et 120 périodes par seconde, mais nous n'en avons enregistré en général qu'une bande comprise entre 30 et 50. La période naturelle du mécanisme transmettant les composants du mouvement est de l'ordre de 0,05 s. et varie selon l'amplification réglable ; le tout permet une lecture généralement aisée des enregistrements.

L'ensemble de ce dispositif permet des avantages considérables sur ceux qu'utilisaient les géophysiciens allemands. Son poids, y compris celui des explosifs, n'excède pas 600 kg, pour une campagne ordinaire d'une durée de 20 jours. Après un essai satisfaisant au glacier du Rhône en 1935, ce matériel a servi à toutes les campagnes de sondage sismique de la C. H. Gl., sans parler de très nombreuses prospections géophysiques indépendantes. Nous nous bornerons à résumer ici les principaux résultats obtenus dans le domaine du sondage glaciaire.

V. — Sondage du Glacier Inférieur de l'Aar

Situé dans les Alpes Bernoises et à la source de l'Aar, ce glacier est l'un des plus grands de la Suisse ; il alimente depuis 1931 le grand lac artificiel du Grimsel pour la production de l'énergie électrique. Par sa longueur totale de plus de 15 km, il occupe le second rang des glaciers suisses et par sa superficie de 39 km², il occupe la quatrième place. Rappelons qu'à la suite des mémorables travaux de *Louis Agassiz* et de ses compagnons, de *Dollfuss-Ausset*, etc. puis récemment des mensurations entreprises par la Compagnie des Forces Motrices de l'Oberhasli, il est un des mieux connus, justifiant le choix de la C. H. Gl.

La première campagne de sondage eut lieu en 1936 dans la partie inférieure (Untere Brandlamm) où la vitesse de propagation des ondes longitudinales se révéla d'emblée un peu supérieure (3700 m/s.) à celle trouvée au glacier du Rhône (3600/m.s.).

Il apparut enfin qu'ici le glacier ne reposait pas directement sur le lit rocheux, mais sur un dépôt fluvio-glaciaire formant moraine de fond.

La seconde campagne, en 1937, s'effectua un peu plus en amont (Obere Brandlamm), où le glacier repose directement sur le lit rocheux. La vitesse de propagation des ondes longitudinales était là de 3800 m/s. dans la glace et de 4750 m/s. dans la roche. La fin de la campagne fut consacrée à la détermination de l'épaisseur des alluvions remplissant le bassin du Räterischboden, à 2 km. en aval de l'Hospice du Grimsel, où on trouve une couche de 40 m environ dans laquelle la vitesse de propagation des ondes longitudinales était de 1470 m/s.

La troisième campagne, en 1938, commença par un sondage sismique des alluvions fluvio-glaciaires et des dépôts morainiques accumulés devant le front du glacier, sur l'Aarboden, alors que le niveau bas des eaux du lac permettait encore de telles opérations. Les vitesses de propagation des ondes longitudinales y furent trouvées assez changeantes, amenant à voir dans ce terrain des alluvions fluvio-glaciaires d'une quarantaine de mètres d'épaisseur au maximum. On confronta ces résultats avec ceux d'une prospection électrique effectuée dans les mêmes lieux par M. W. Fisch (Zurich). Malgré les difficultés rencontrées par ce dernier, qui soupçonna la présence de « glace fossile » à l'intérieur des dépôts, la concordance s'établit à quelque 20 % près.

La campagne se poursuivit sur le glacier même, dans les parages de la cabane de Lauteraar (Pavillon Dollfus) confirmant que la réflexion des ondes était bien meilleure sur le lit rocheux, que le glacier recouvre dans ces parages. La précision des résultats fut dès lors excellente. Malgré une profondeur de 300 m environ, les profils longitudinaux et transversaux se correspondent à 3 à 4 m près, soit 1 %.

En 1939, une quatrième campagne permit de sonder de plus grandes profondeurs dans la région de l'Abschwung où se réunissent les deux affluents principaux du glacier, soit ceux du Lauteraar et de Finsteraar. La vitesse de propagation des ondes longitudinales atteignit là 3850 m/s. dans la glace, en relation semble-t-il, avec la profondeur qui mesure dans ces parages près de 450 m. Cette augmentation de vitesse n'est d'ailleurs pas encore interprétée complètement, mais il faut sans doute la rechercher dans la pression plus grande de la glace profonde. La vitesse des ondes transversales est de l'ordre de 1700 m/s. soit de 2,1 à 2,3 fois plus petite; leur amplitude est très grande, mais leur arrivée au sismographe n'est pas aussi nette que celle des ondes longitudinales et on ne les a guère employées; enfin, les ondes superficielles ont été enregistrées aussi, mais pas utilisées.

En 1947, après une longue interruption due à la guerre mondiale, une dernière campagne a permis de prospecter les deux affluents en amont de l'Abschwung, révélant une disparité très nette de leurs lits, le plus profond étant celui du Finsteraar où l'épaisseur du glacier semble dépasser 500 m (fig. 2). Les résultats définitifs ne sont pas encore connus.

En 1948, un sondage complémentaire doit permettre de compléter la connaissance du lit dans la région comprise entre l'Obere et l'Untere Brandlamm, ainsi que de fixer la limite entre la moraine sous-glaciaire et le lit rocheux en contact direct avec le glacier.

Au total, on a obtenu et effectué sur ce glacier 750 sondages, permettant d'établir une carte sommaire du lit portant des points cotés (et, pour certains, les valeurs de la pente), des profils transversaux et un profil longitudinal de la partie inférieure du glacier, entre le portail et l'Abschwung, longue de 7 km. Dans cette région terminale la surface du glacier a une largeur moyenne de 1 km et descend dès l'Abschwung avec une pente moyenne de 8 %, assez régulière. Le lit a une forme différente, moins régulière et non dépourvue de contre-pentes, telles qu'on les connaît bien dans les terrains abandonnés par les glaciers, riches en cuvettes. Abstraction faite de l'arête rocheuse qui plonge de l'Abschwung entre les deux affluents avec une pente de 66 %, la pente moyenne du lit est très faible, soit de 2 % seulement dans sa partie terminale, sur 6,5 km. Aussi l'épaisseur du glacier croît-elle rapidement du portail en amont, ainsi que le montre le tableau suivant :

A	0 km du portail	épaisseur de	0 m
	1 km		148 m
	2 km		190 m
	3 km		300 m
	4 km		350 m
	5 km		365 m
	5,75 km		430 m (maximum)
	6,5 km		360 m (près de l'Abschwung)

(Les épaisseurs indiquées sont rapportées à la surface glaciaire en 1936-39, dont les niveaux se sont abaissés depuis lors de 10 à 15 m, selon les régions. Fig. 3.)

VI. — Sondage du Glacier de Morteratsch

En 1943, la C. H. Gl., que les circonstances nées de la guerre empêchaient de poursuivre les travaux au Glacier Inférieur de l'Aar, réussit cependant à faire une campagne sur le glacier de Morteratsch, le plus important du massif de la Bernina (Grisons, 21 km² de superficie et 8 km de longueur environ). choisi pour son accès facile. La campagne fut courte mais riche d'enseignements : on obtint 118 sondages, dont 100 sur le glacier et les autres dans la moraine déposée en aval. Il s'avéra aussi possible d'utiliser des microphones pour l'enregistrement des ondes transversales. Le dépôt glaciaire devant le glacier se révéla peu important (10 m environ), et la vitesse de propagation des ondes longitudinales dans ce matériel a été trouvée égale à 1600 m/s. Dans la glace on trouve 3650 m/s. S'agissant d'une détermination effectuée dans la partie terminale du glacier, elle concorde bien avec les résultats trouvés au Glacier Inférieur de l'Aar. Quant à l'épaisseur du glacier même, elle n'est que de 70 à 100 m à 1 km environ en amont du portail, mais à une distance de quelque 4,5 km elle atteint la valeur de 360 m., considérable pour un glacier de cet ordre. Dans cette région, située à 1 km au sud de la cabane Boval du C.A.S., le bassin rocheux sous-glaciaire offre une contre-pente nettement accusée en aval, avant la confluence du Vadret da Morteratsch avec celui de Pers.

VII. — Sondage du Glacier de la Plaine Morte

Au cours de l'été 1946, la C. H. Gl., mettant ses instruments à la disposition de M. Kreis, lui permit une expertise demandée par une corporation de droit public qui se propose de capter de l'eau sous les glaciers de la Plaine Morte et du Wildstrubel (Oberland Bernois). Il s'agit d'un plateau glaciaire qui s'étend sur 7 km environ de l'Ouest vers l'Est, et sur 4 km environ du Sud au Nord où il forme le glacier de Rätzli dont les eaux sont tributaires de l'Aar. De cette vaste étendue, seule la partie centrale a été prospectée, mais elle est aussi la plus intéressante. Située à l'altitude moyenne de 2800 m, elle reste en été recouverte de neige, ce qui a obligé les sondeurs à enfoncer les charges explosives à 2 m au lieu de 1 m et à en augmenter les doses. Les réflexions sur le fond furent excellentes et donnèrent une profondeur moyenne de 200 m (maximum 225 m). C'est une valeur assez considérable pour un glacier aussi petit, dont le bassin d'accumulation ne s'élève guère au-dessus de 3000 m. Le lit du glacier de Rätzli même atteint probablement une profondeur plus grande encore, ainsi qu'il semble résulter du sondage incomplet effectué vers l'aval où la profondeur atteint déjà 245 m. Les travaux de captage d'eaux n'ayant pas encore été entrepris, cette configuration probable du lit n'a pu encore être vérifiée.

VIII. — Sondage du Gletschboden au Glacier du Rhône

Effectué à la fin de l'été 1946, ce sondage avait pour but de déterminer approximativement la puissance des dépôts fluvio-glaciaires et morainiques qui remplissent le vallon situé entre la station de Gletsch et la bas de la cataracte du glacier du Rhône, dépôts qui constituent le Gletschboden d'où le glacier s'est retiré progressivement depuis plus d'un siècle. Il n'y a pas lieu d'exposer ici la méthode utilisée pour cette prospection dans laquelle la détermination du lit rocheux s'est faite essentiellement par l'interprétation des hodographes des ondes P en partie réfractées par les versants rocheux. Dans l'alluvion la vitesse de ces ondes atteint la valeur élevée de 1820 m/s en raison de la forte teneur en eau du terrain, ce qui a permis du reste de se borner à faire exploser les charges à la surface ou dans l'eau et d'économiser ainsi le forage coûteux des trous d'enfouissement. On a admis pour la vitesse dans la roche encaissante la valeur de 5000 m/s (+ ou - 200 m/s). Une coupe transversale établie à 1 km environ en amont de la Chapelle de Gletsch donne au lit rocheux un profil en cuvette allongée, large de 475 m et d'une profondeur maximale de 120 m.

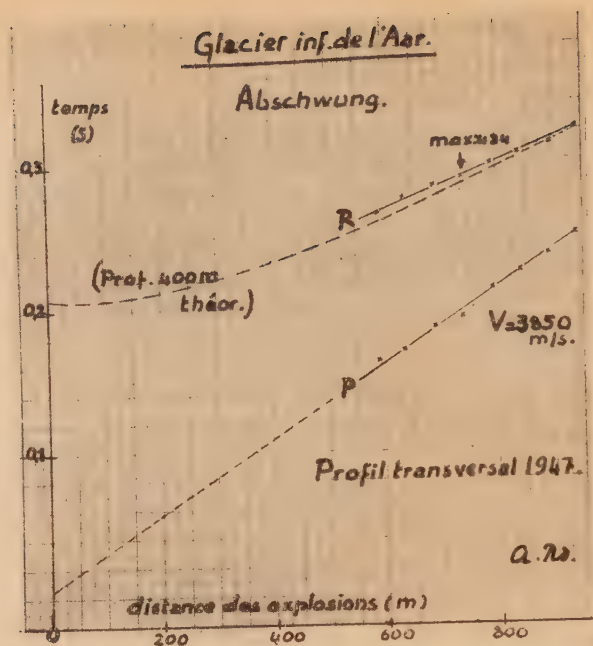


FIG. 2

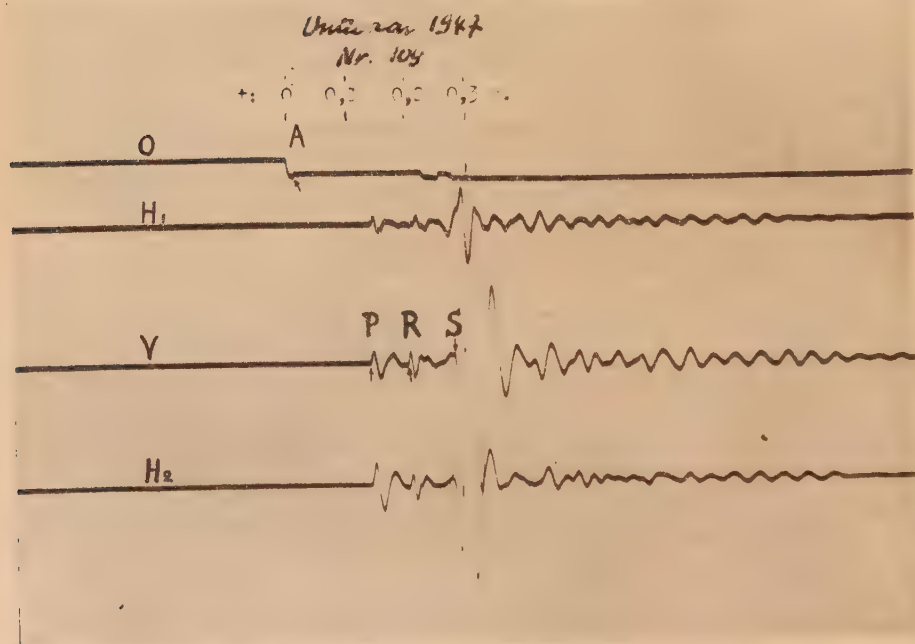


FIG. 1

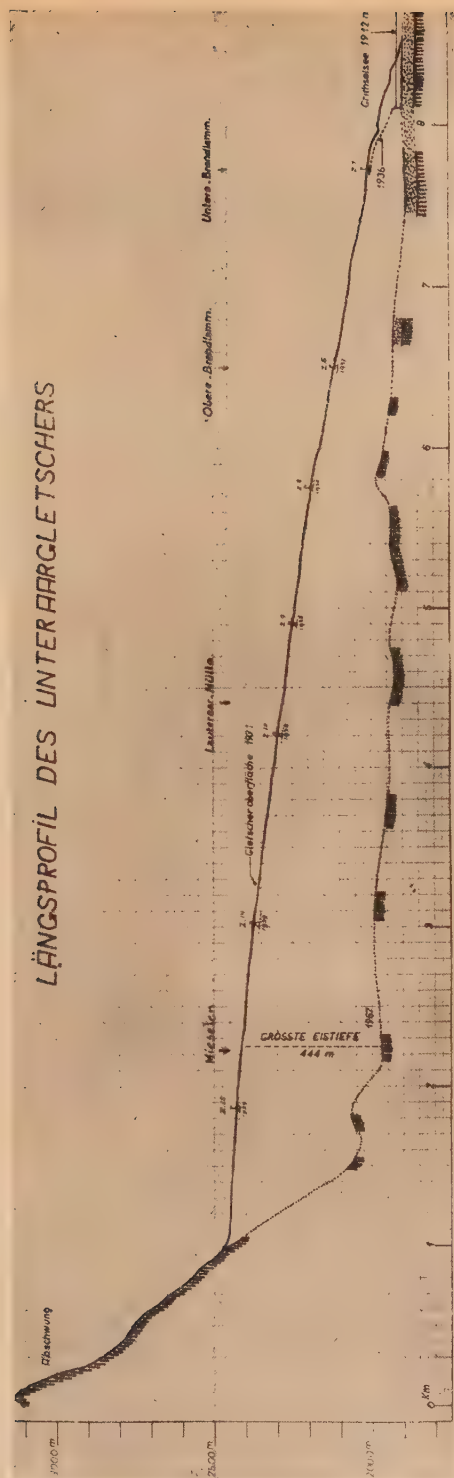


FIG. 3

IX. — Sondage du Grand Glacier d'Aletsch

En 1947, la C. H. Gl. mit ses instruments et une partie de ses ressources financières à la disposition de M. *Haefeli* (Zurich) pour lui permettre de sonder ce glacier quelque peu en aval de Concordiaplatz, région très peu profonde sondée en 1929 par les géophysiciens de Göttingen qui y trouvèrent une épaisseur dépassant 700 m. Les résultats du dernier sondage sont consignés dans le mémoire présenté d'autre part par M. *Haefeli*. Nous nous bornerons à relever ici que la profondeur trouvée est très inférieure aux premières, révélant donc l'existence d'une contre-pente en aval de la cuvette de Concordiaplatz. D'autre part, les sismogrammes n'ont pas fourni de réflexions très nettes et ceci est en relation, sans doute, avec un fond rocheux vallonné ou quelque revêtement morainique.

X. — Conclusions

Ce résumé des sondages sismiques de la Commission Helvétique des Glaciers (C. H. Gl.) montre que cette méthode de prospection a été appliquée avec succès à plusieurs glaciers suisses. La C. H. Gl. se propose de persévérer, notamment au glacier d'Aletsch, en particulier dans la région du névé, au voisinage de la Station de Recherches Scientifiques du Jungfrauoch.

André RENAUD,
(membre de la C. H. Gl.)

Lausanne, juillet 1948.

BIBLIOGRAPHIE

1. — P. Ls. MERCANTON : Les variations périodiques des glaciers des Alpes suisses. (Rapports annuels N^{os} 47 et suivants, parus dans « *Les Alpes* », revue du Club Alpin Suisse dès 1927).
2. — P. Ls. MERCANTON : Les sondages sismométriques de la Commission Helvétique des Glaciers. In *Actes, de la Société Helvétique des Sciences Naturelles*. Soleure 1936.
3. — W. JOST : Die seismischen Eisdickenmessungen am Rhönegletscher 1931. (*Mémoires de la Société Helvétique des Sciences Naturelles*, Vol. LXXI. Mem. 2, 1936).
4. — A. KREIS : Ergebnisse der seismischen Eisdickenmessungen auf dem Unteraargletscher, *Verh. des Schweiz. Naturforsch. Gesellschaft*, Basel, 1941.
5. — R. HAEFELI : *Observations in the Firn and Ablation Region of the Great Aletsch Glacier*. Réunion de l'U.G.G.I. Oslo, 1948.

12) PRELIMINARY GLACIOLOGICAL PLANS FOR THE NORWEGIAN-BRITISH-SWEDISH ANTARCTIC EXPEDITION, 1949-52

by

Wilson AHLMANN (Stockholm)

An expedition to Dronning Maud Land in the Antarctic sector of the Antarctic — chiefly for the purpose of glaciological, meteorological and geological studies — was first proposed in London and Oslo in May 1946. This aroused great interest in the three countries — Norway, Great Britain and Sweden — whose participation was suggested. The Norwegian Polar Institute (Norsk Polarinstitutt) had been founded in March 1948, and Professor H.U. Sverdrup, returning from the United States in May to be its Director, was appointed chairman of the Norwegian Antarctic Committee. In July the *Storting* granted sufficient funds for the requisite preparations. The Norwegian Government sent special notes to the Swedish and British Foreign Offices, inviting their respective countries to participate in the expedition (1).

The expedition will sail under the Norwegian flag and the leadership of Professor Sverdrup. The winter party will consist of thirteen or fourteen members, of whom six should be specially qualified scientists, two from each country.

The expedition will charter a modern Norwegian sealing vessel for transport, but it must also acknowledge its gratitude to the Norwegian whaling companies for undertaking to carry the heavier equipment including aircraft. In early January 1950 we hope to reach Dronning Maud Land close to the meridian of Greenwich where the shelf ice border is low.

It is planned to spend two winters in the Antarctic. The chartered ship will visit the wintering party in the summer of 1950-51, and the expedition will leave the Antarctic in 1952. Dog teams and weasels will be used for land excursions; the aircraft, either a helicopter, an aeroplane, or both, will undertake the longer journeys.

Some parts of the interior of Dronning Maud Land were surveyed from the air by the German *Schwabenland* Expedition in 1938-39. The resulting photographs and reports, published in 1942, indicate that these districts are extremely interesting, especially from the glaciological point of view.

The so-called « Wohlthat-massif » of the area which the Germans called « Neu-Schwabenland », about 10,000 sq. km. in extent, is surrounded and partly covered by inland ice the surface of which is 1000 m. above sea-level in the north and 3000 m. in the south. In that as well as in other *massifs*, extensive areas free of ice and snow form « oases » in the ice desert, with local glaciers of various sizes and morphological types. The detailed German maps also show some lakes, the largest of which — 650 m. above sea-level — is 6 km. long and 4 km. broad. Smaller lakes, are marked on the maps up to 850 m. above sea-level. It is very difficult to imagine the existence of liquid water at these heights and so far inland in present climatic conditions.

About 250 m. above sea-level, however, there are small lakes and ponds of liquid water both on the shelf ice and on the low-lying land projecting above it. Terminal and lateral moraines prove that the lobes of the inland ice in the broad spaces between the mountain ridges were formerly much thicker and larger than they are now. Small terminal moraines in front of the present local glaciers in these « oases », as well as long moraines at the bases of the mountain ridges, must have been left by earlier glaciers larger than the present ones. According to R. van Klebelsberg the topography of the ground in front of the glaciers in the area which the Germans called « Kurze Gebirge » indicates dead ice. In these relatively limited areas we have accordingly unmistakable proofs of a former greater extension of the ice, and probably also of a considerably milder period. The rates of movement, and possibly also the regimes of the local glaciers, should be investigated. Thanks to a large number of nunataks of various sizes and positions, it will also be

(1) The Swedish contribution was approved by the *Riksdag* in November 1948.

possible to determine the rate of movement of the inland ice surrounding and intersecting the whole district in streams of varying widths and gradients.

The photographic material obtained by the German expedition was abundant and of the highest quality. A small but instructive part of it was published in 1942, but most of it was destroyed, or at least is not at present available in western Germany. Some photographs have been saved at Hamburg, however, and will be placed at our disposal. We hope that what is left will, on comparison with new photographs taken from the air by our expedition, allow us to draw some conclusions regarding changes in the extension and thickness of the ice in the last ten or twelve years.

The glaciological work of the expedition will be done (1) at headquarters, (2) by flights inland to the mountain district and (3) by excursions with dog sledges or mechanical transport in that district.

In the preliminary glaciological programme for the expedition drawn up by W. L. S. Fleming and myself at a meeting in Oslo in July 1947, we pointed out that in measuring the accumulation special attention must be paid to the difference between precipitation and snow drift. As regards the latter, Bagnold's methods for measuring sand drift at different altitudes above the ground may prove the most practical. Direct precipitation might possibly be distinguished from drift-accumulation by studying the form of the snowflakes or grains in each case. Observations of ablation should be made with a view to determining how much of the ablation can be ascribed to melting, to evaporation and to wind erosion (corrasion). An important but very difficult problem is to devise a method of differentiating between the ablation attributable to evaporation and that due to wind erosion.

The ablation measurements are intimately associated with the whole problem of the heat economy of the snow or ice surface — the balance between incoming radiation, conduction, condensation, reflection and outgoing radiation. During the Norwegian-Swedish expedition to West Spitsbergen (Vest Spitsbergen) in 1934 professor Sverdrup devised the method of calculating the heat economy of glaciers, and C. C. Wallén has since continued these investigations on a small glacier in the far north of Sweden. The relation between radiation and conduction in the ablation processes varies with latitude and altitude as well as with the progress of the ablation season. We know little or nothing of these conditions on the Antarctic continent.

The most important task at the base station will be to study the physical properties of the inland ice, which is of the high polar type, *i.e.* its temperature is negative down to a considerable depth.

The geophysical and meteorological conditions in the Antarctic differ so much from those in the Arctic — except in the central and northernmost part of Greenland — that the results of the more detailed investigations round the North Atlantic coasts cannot be directly applied in the Antarctic. While all glaciers in the northern hemisphere have typical accumulation and ablation areas, the Antarctic inland ice is a single, almost continuous, accumulation area, the equilibrium of which is maintained by calving. Particularly on account of the low summer temperature, the ice masses of the Antarctic must presumably differ essentially in internal structure and movement from those of the northern hemisphere. The mechanical influence of the Antarctic glaciers on their substrata is, moreover, likely to differ from what is known of temperate glaciers.

Much evidence indicates that appreciable climatic variations occurred in northern Europe during and after the recession of the last Pleistocene inland ice, and it is probable that these variations were more or less concurrent in large parts of the globe. Very little is known of conditions in the Antarctic, but similar variations may well have occurred there.

We are planning to record by electric resistance thermometers the temperature of the frozen firn and ice at different depths down to about 100 m. A thorough study of the temperatures hitherto observed shows that no one has so far reached a sufficient depth to be able to answer all the questions necessary for a full understanding of polar glacier physics. We have, for instance, as yet no certain basis for calculating the depths at which melting point is reached at any of the places where temperature observations have been taken. Of Borg, the headquarters of the expedition of J.P. Koch and Alfred Wegener to Greenland in 1912-13, situated 62 m. above sea-level in lat. $76^{\circ} 41' N.$ and long. $22^{\circ} 24' W.$, Wegener only states that it is at any rate proved that to a depth of at least 140 m. the glacier temperature must be negative. No light is thrown on these problems from the famous «Eismitte» station of Wegener's expedition of 1930-31, 3000 m. above sea-level nor from the «Weststation» of the same expedition. Less than 6-7 m. below the

surface the seasonal variations were very marked. Deeper down the mean annual temperature first rose very slowly to a depth of about 12 m., after which it decreased slowly at the rate of 0.04° C. per metre. This is remarkable, considering that the temperature in the deeper sections of the inland ice must presumably be raised to near melting point by the transfer of heat from the earth. E. Sorge suggests that in this case the fall of the temperature with increasing depth may be due to the fact that the air temperature on the inland ice, as elsewhere in Greenland, has in recent years been higher than before.

The U.S. Antarctic Service Expedition of 1939-41 took temperature observations on the Ross shelf ice down to a depth of 41 m., where the temperature was nearly constant at -22° C. According to the seismic sounding taken by the United States Operation «Highjump» 1946-47, the thickness of the Ross shelf ice was only 200 m.

The possibility of differentiating between the several seasonal layers is also of great importance to all problems connected with the physical properties, structure and regime of all glaciers, local as well as inland ice. This problem can be investigated in open pits and by cores drilled from undisturbed material at different depths. Koch and Wegener came to the conclusion in Greenland in 1912-13 that the previous winter's layer could be distinguished from the new layer below it, but that the rest of the firm deeper down in the inland ice was not stratified at all. At «Eismitte» on the other hand, they were of the opinion that seasonal layers could be traced with certainty right to the bottom of their shaft, 16 m. deep. At the top the seasonal layers could be seen with the naked eye; deeper down they could be recognized by the size of the firm grains, which are somewhat larger when deposited in the summer than in the winter. Below a depth of 6 m. the varying densities of the frozen firm provided evidence which was considered satisfactory. No annual stratification is believed to occur in the Antarctic. Its existence has actually been definitely denied, but I hesitate to accept that dictum as correct. All the methods used at «Eismitte», and every other conceivable means, must at least be tried before giving up hope completely.

The last but not the least important item of the base station programme is the crystallographic investigation of the inland ice. It is highly desirable to follow the structure of the frozen firm and the ice as far down as possible. We do not know at what depth the frozen firm turns into true ice, but we will try to reach a depth of about 150 m. This necessitates the drilling of undisturbed cores by a specially designed machine. Technically, the most difficult problem is to remove the cuttings from the bottom of the hole. Our experts on drilling machines recommend alcohol as a flushing medium and do not believe that this would affect the structure of the core.

It will be important, especially to the glaciologists, to take as many photographs as possible of the «Neu Schwabenland» district for comparison with those taken by the Germans. If practicable one or more local glaciers should also be selected for detailed study.

The most important object of these regional studies is to ascertain whether the recent climatic fluctuation has or has not extended to the Antarctic. We know that for one or more decades most glaciers in the world have been shrinking at an increasing rate. Those round the northernmost Atlantic, and in the whole Arctic region, are receding particularly rapidly owing to a climatic improvement which takes the form of higher temperatures especially in winter, spring, and autumn. The ablation season has been lengthened to such a degree that the total amount of ablation more than compensates for the small increase in the precipitation of snow and hoar frost. One speaks of increased atmospheric circulation accelerating the transfer of heat from low to high latitudes. The present climatic fluctuation is undoubtedly the most remarkable that has occurred for the last two hundred years, and is moreover a phenomenon of great economic significance, not least to the countries of the north. It is accordingly attracting the increasing interest of meteorologists, glaciologists and biologists. The problem cannot be discussed in its entirety until we know whether this climatic fluctuation extends also to the Antarctic or not. If it does, the climatic change is global and its cause probably must be sought outside the earth. If it does not, the perspective alters.

One consequence of the present glacier shrinkage is the rise in the levels of the oceans, calculated at about 1 mm. per annum. This relatively small figure indicates that the Antarctic and the bulk of the Greenland inland ice — which between them contain more than 90 per cent of all the present-day ice covers — cannot have diminished at the same rate as the local glaciers. If the polar inland ice sheets

should begin to melt as rapidly as the other glaciers, the rising sea-level would become a phenomenon of great far-reaching practical importance. The whole problem of eustatic movements — the rise and fall of the sea surfaces — must remain more or less unsolved as long as we do not know for certain whether, and to what extent, the Antarctic has contributed to these volumetric changes.

Every effort will also be made to record the meteorological and climatological observations necessary for a proper understanding of glaciological processes.

VII. — Rapports relatifs à la :
FORMATION DE LA GLACE
SUR LES LACS ET LES RIVIÈRES



1) ICE FORMATION IN LAKES AND RIVERS

by

OLAF DEVIK

In a paper in the *Geofysiske Publikasjoner*, vol. IX, No. 1, 1932, I have treated quantitatively the different factors in the process of heat exchange at a water surface in open air. By the formulae and tables given in that paper, it would be possible to calculate the net loss of heat from a water surface under given meteorological conditions. Corresponding tables also give the net loss from ice surfaces or from snow surfaces. I will not enter into detail, but very briefly quote some figures to give an idea of the loss of heat and the corresponding growth of ice under some typical meteorological conditions. When I escaped from Norway in March 1943 I was not able to bring any papers with me, and consequently the results from investigations in recent years have had to be quoted in the present paper only from memory. For the same reason no complete list of my publications on this matter in recent years could be included.

Consider first the sunlight and the daylight radiation in northern latitudes in winter. At a low sun-altitude, say $5-10^\circ$, a water surface will reflect a considerable part of the sun rays. A fraction will enter the water and be gradually absorbed, the transmitted radiation being practically all absorbed in the bottom of the lake or river. The unit of heat is the calorie, the amount of heat which is needed to raise the temperature of 1 gramme of water by 1°C . At a sun-altitude of 5° , 1 square centimetre of water surface will receive only 0.6 calorie per square centimetre per hour. This will be only 0.5 per cent of the amount of sun radiation reaching the atmosphere.

In these conditions and with clear sky the diffused daylight from all parts of the sky will give a greater input, namely, 3 cal./cm²/hour. If the sky is clouded, the diffused daylight absorbed in a water surfaces will be less, but still represent 1 cal./cm²/hour, which is more than the direct radiation would give at the low sun-altitude considered.

These figures of heat gained are small compared with heat losses in the winter. But when the sun's altitude increases, the incoming radiation, direct and diffused, will contribute more and more to the heat absorbed; and in the summer the relation will be the other way round, the heat given will exceed the heat lost.

Confining ourselves in this paper however to the conditions in mid-winter, say at Christmas time, we then pass to the loss of heat by infra-red radiation and choose as a representative condition an air temperature of -10°C . With a clear sky the heat radiation from a water surface at 0°C . will be 13.8 cal./cm²/hour. In the case of a clouded sky, the amount is much less, 4.9 cal./cm²/hour.

It is a very important fact that practically all non-metallic surfaces will absorb almost completely the infra-red radiation of the type radiated from objects of moderate temperature. Thus the radiation from a hot kettle will be absorbed completely in the surface layer of water which, on the other hand, is transparent to ordinary visible rays of light. From the infra-red point of view, water is a «black body». This is still more the case with snow. The perfect «white» surface of snow relates only to visible light: to infra-red radiation of the type we mentioned snow is a perfect «black body». This means that infra-red radiation falling upon a water surface or a snow surface will be completely absorbed in the uppermost layers, in a very thin sheet of some hundredths of a millimetre thick. On the other hand, the same layer will emit infra-red radiation of the same type to the atmosphere.

The next process we have to consider is the loss of heat by convection. This loss will be due to the difference in temperature between the water surface and the air above, but it will also be modified by the wind. Everybody knows that a vessel or a body is much more rapidly cooled down when a cold wind is blowing than when it is calm. To give the figures relating to a water surface: with an air temperature of -10°C . and calm, the heat loss by convection will be 2.8 cal./cm²/hour. With a wind velocity of 5 metres per second, the loss will be 11.5 cal./cm²/hour.

There is still one loss which we have not yet considered — that due to evaporation. The evaporation will depend first, upon the dryness of the air above the water surface or, more exactly, upon the vapour pressure, and secondly, upon the wind. With an air temperature of $-10^{\circ}\text{C}.$, a vapour pressure of 2.6 mm of mercury, and calm air, the loss by evaporation from a water surface will be 1.7 cal./ cm^2 /hour, but if there is a wind of 5 metres per second, it will increase to 7.7 cal./ cm^2 /hour. It is an everyday experience when bathing in open air that the cooling caused by wind is very strong when the skin is still wet.

Next, if we add all the heat losses in water and subtract the gain of heat, the net heat loss under the first set of meteorological conditions, air temperature $-10^{\circ}\text{C}.$, vapour pressure 2.6 mm, wind velocity 0 m./sec., cloudiness 0, sun altitude 5° , water temperature $0^{\circ}\text{C}.$, will be 14.7; and in the second set, air temperature $-10^{\circ}\text{C}.$, vapour pressure 2.6 mm., wind velocity 5 m./sec., cloudiness 10 (overcast), sun-altitude 5° , water temperature $0^{\circ}\text{C}.$, it will be 23.1 cal./ cm^2 /hour. These figures are equivalent to a production per hour of a sheet of ice of thickness 2.0 and 3.2 mm respectively.

The growth of a sheet of ice

The analysis which we have briefly outlined in the preceding paragraph forms the basis for the calculation of the actual growth of ice on still water. We could simply apply the figures directly if the surface of the ice retained the temperature of $0^{\circ}\text{C}.$, and had the same physical properties as water with regard to heat exchange. The first supposition is however entirely wrong because the surface temperature of the ice will be the lower the thicker the ice, supposing the air temperature is low.

Fig. 1 is a diagram of the temperature distribution in an ice layer of growing thickness. In A the sheet of ice is very thin and the temperature gradient is very

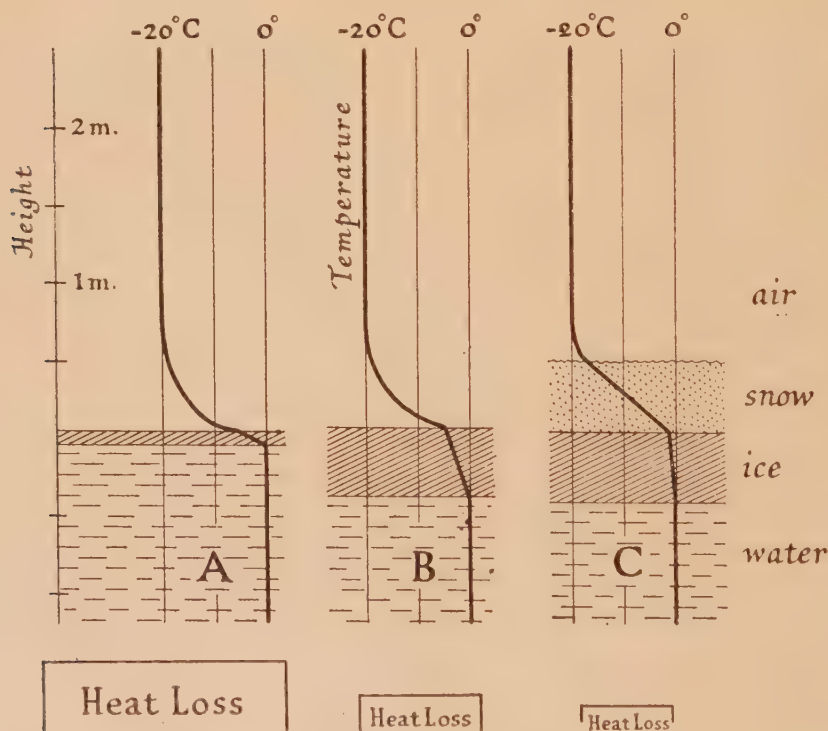


FIG. 1. Temperature distribution in A, water, thin ice, and air; B, water, thick ice, and air; C, water, thick ice, snow cover, and air.

steep. The loss of heat is correspondingly great and the ice-thickness will increase rapidly. In B the ice had attained considerable thickness, the temperature gradient in the ice is much smaller than in A, and the loss of heat correspondingly less. In the third case, C, the ice layer is covered by a considerable layer of snow with the result that the temperature gradient in the ice is very small and the heat flow is correspondingly reduced. The gradient in the snow is considerable but the thermal conductivity is much smaller than in the ice. The net result is that the surface of the snow will have a very low temperature, practically the same as that of the air. The resulting heat loss will be very small.

If we turn the diagram through 90° it will also represent the temperature distribution under winter conditions through the outside wall of a house of varying thickness and with an isolating layer. This may be of some interest when we are pondering on the insulating qualities of prefabricated houses.

Ice formations and supercooling of the water surface

The crystallization process does not start by itself. It requires starting nuclei, *i.e.* particles of solid matter which have already a crystalline structure. In open water these particles will always be present; it requires a high degree of precaution to clean water from all sorts of possible nuclei.

When an ice crystal has started growing upon a solid particle, the heat of solidification, 80 calories per gramme of water, must be carried away, which means that the temperature outside, or inside, the crystal surface must be lower than the temperature at the crystal surface itself, which is 0°C . Therefore the water in which an ice crystal is growing will generally have a temperature below zero or be supercooled. To measure this supercooling is however not always an easy matter.

We may use a precision thermometer with very fine divisions. But if the thermometer is brought from the cold air into the open water the bulb will immediately be covered by a thin sheet of ice and the measurement is spoilt, the thermometer only indicating the freezing point. In order to perform the measurement we must heat the bulb above 0°C . and then immerse it in the water. With proper precautions it is not difficult to observe supercooling effects of some hundredths of 1°C . The freezing point on the thermometer is best observed simply by covering the submerged thermometer with ice formed on the spot. The measurement will however only give an average of the temperature in a water layer of the same thickness as the size of the bulb. If we want to measure the temperature of the surface layer itself where the heat exchange with the air actually takes place we must adopt other methods.

During the winter of 1941-42 I performed a series of measurements with a Moll thermopile placed in a special mounting and provided with photographic recording of the potential delivered by the thermopile. Fig. 2 is a sketch of the arrangement. I used a Moll thermopile of the large type with 80 thermoelements placed within a receiving area of diameter 20 mm. The chief difficulties arise from the fact that the temperature will be very unevenly distributed in the space where the thermopile is to be placed: at the water surface the temperature will be near 0°C ., but it will decrease rapidly with height, till the general air temperature is reached. Moreover this temperature distribution will change with every puff of wind. The thermopile is extremely sensitive to temperature differences in the adjoining metal housing and it is essential to reduce the temperature gradients in the material of the cone and the diaphragms of the free opening and to keep all changes of these gradients slow.

Both requirements are met by using a casing with a great mass of good conducting material, preferably copper or brass, by machining the diaphragms out of the material itself, and by using a motor ventilator to obtain evenly distributed temperature of the housing just before the measurements commence.

The thermopile being exposed to the radiation from the surface, the temperature of which is to be measured, will produce a voltage which is the greater, the greater the difference of temperature between the thermopile and the surface. In our case the water surface will have a temperature of nearly 0°C . while the air at the place of the thermopile casing will be much colder, for instance -15°C . We are however not interested in measuring this temperature difference, but in the changes which are due to changes of the temperature of the radiating water surface. The

voltage which is produced by the large mean temperature difference is therefore compensated, and only the changes are measured by the galvanometer (see Fig. 2).

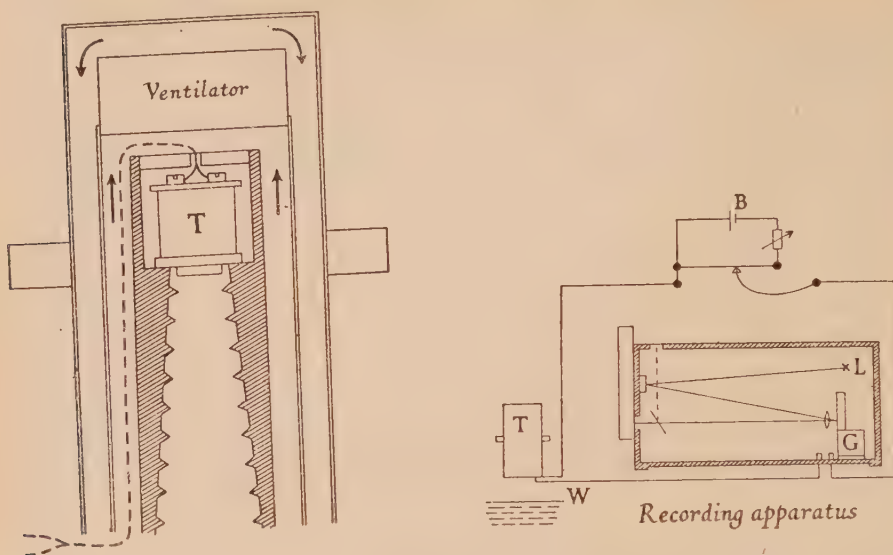


FIG. 2 Thermophile in housing and recording apparatus. T, thermophile; G, galvanometer; L, lamp; B, compensating bridge; W, surface of water.

The measurement is done by photographically recording the movements of the light spot of the galvanometer. The calibration of the recorded curve is done by using the radiation from the surface of rapidly running water taken alternately from two sources of known but different temperature.

To observe the surface temperature of water exposed to the free air in order to measure the supercooling effect it is best to avoid the sun-radiation and perform the measurements before sunrise. Having reestablished one of my observation stations in a mountain region near Rondane (at Soernesset) shortly after Easter 1942, I have never in my life enjoyed so many brilliant sunrises terminating the preceding hours of work.

In one series of experiments I used an open basin with inlet and outlet for the water, the temperature of which was slightly above 0°C . I allowed a sheet of ice to be formed and then suddenly increased the flow of water, causing the ice to be covered by a thin layer of water, starting with a temperature slightly above zero. I placed the thermopile above the basin at a distance of about 10 cm with the opening downwards. The typical result of such measurements is given in Fig. 3.

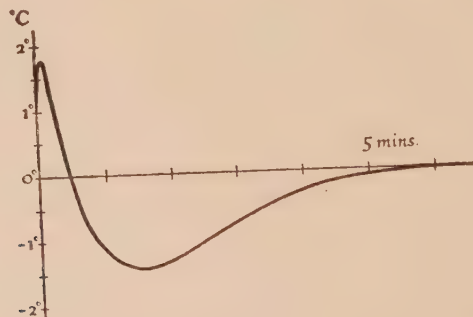


FIG. 3. Surface temperature of freezing water recorded by thermopile

When the decreasing temperature of the water surface passes 0°C ., nothing happens, but the temperature falls farther and within a comparatively short time, between one or two minutes in this case, a temperature of approximately -1.5°C . is reached. Then the formation of ice crystals starts, needles grow over the surface from starting points, and in the course of the next minutes a network is formed growing denser and denser. The temperature registered by the thermopile rises correspondingly. This is because the temperature of the open spaces are still considerably below zero while the temperature of the thin needles will be practically zero. In fact if we drop some ice powder in the open spaces between the needles these will immediately be covered by a leaf of ice, indicating that the water had been considerably supercooled. When, in this case, five minutes had passed a coherent sheet of ice had been formed.

Fig. 4 shows how the growth of ice takes place when an ice crystal is starting to grow. The dimensions are exaggerated and the conditions simplified. The arrows indicate the growth, which will of course be greatest towards those places

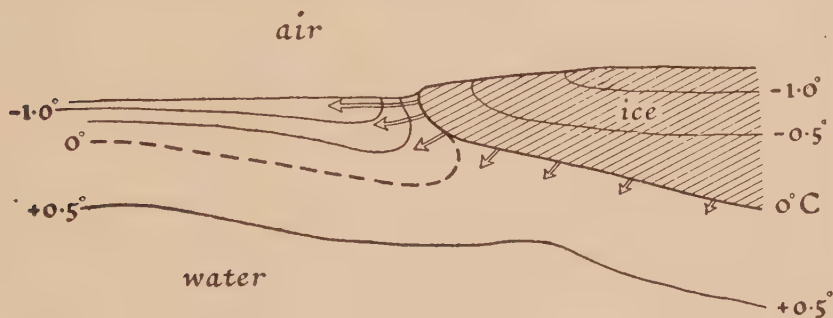


FIG. 4. Growth of an ice-crystal in supercooled water.

where the temperature gradient is steepest. From the temperature distribution in the neighbourhood of the ice crystal surface the growth will evidently be greatest just in the surface layer. That is the reason why ice needles can shoot over the surface in an amazingly short time.

It is of fundamental importance to the understanding of the ice formation processes to bear in mind that this growth of ice into surrounding supercooled water is a process which takes time. Generally, this fact is not sufficiently realized and I believe that one of the reasons is due to the ordinary presentation of the supercooling effect, for instance, by cooling a solution of thiosulphate and then adding a few small crystals to start the solidification. In this case the crystallization takes place rapidly, but this is due to a very large amount of supercooling and to the comparatively small heat of solidification which is freed by the crystal process. In the case of supercooled water exposed to cooling in open air, the drop of temperature below zero will be small especially when the water is being mixed; and the heat of solidification is very considerable. In this case the process of solidification may take a very long time and we may expect the surprising result, in a lake or a river, of ice particles swimming or floating in supercooled water masses.

I have already mentioned that it is comparatively easy to measure supercooling in cold weather in open rivers which are cooled down about 0°C . The first measurements of supercooling in rivers and lakes were made by Russian observers among whom I might mention particularly W. Altberg. In the running and turbulent water of the Neva he made observations during the years 1916-1917 and 1920-21. He measured supercoolings which generally amounted to some hundredths of a degree C. but in peculiar cases, even -0.1°C . This supercooling was not only measured at the surface itself, but also in the intermediate layers and even at the bottom. He observed the formation of small ice crystals within the water and on the surface of objects under water. The formation of these ice particles did not do away with the supercooling which continued to exist irrespective of the ice crystals and the continuous formation of new portions. This process we shall consider in more detail in the next paragraph.

Ice formation in running water

The figures we have given on p. 183 concerning the loss of heat from a water surface exposed to open air show that under heavy frost the formation of ice would be very rapid as long as this loss of heat continued. In still water however the formation of a coherent ice sheet, a plate which increases in thickness, will more and more effectively reduce the loss of heat. From this point of view the process of ice formation may be considered as a counteraction aiming at reducing the heat loss and establishing a state of stability. This evidently requires a lowering of the temperature of the surface exposed to air approximately to that of the air itself. This is most easily attained when the ice is covered by snow which acts as a most effective insulating layer. (Compare Fig. 1)

In the case of turbulent running water however the conditions are entirely different from those of still water. The water surface will be renewed every moment and will be exposed to the maximum loss of heat, often for very long periods. When the river has been cooled down through the entire mass to about 0°C ., the loss of heat from the surface will produce supercooling. The supercooled sheet of water will however not be resting at the surface, but will be twisted and move in an irregular way through the section of the river. It will enter the turbulent motion and may just as often sweep the bottom as be moving at the surface.

The result of this intermixing will be a changing structure of water masses at different temperatures slightly under 0°C . Wherever such supercooled water currents contain solid particles or pass solid objects, ice will be formed. It is evident that the formation of ice on fixed objects will take place at the highest rate where the motion of the supercooled water passing by is greatest, because there the actual gradient of temperature will have extreme values. The consequence is that every peak of a fixed ice structure will have a tendency to grow, so that the irregularities will not be smoothed out, but on the contrary, exaggerated. The ice which is thus deposited under water is very loose and spongy. The supercooled water will however filter through and gradually fill up parts of the space between the walls and needles of ice. If we put a pole in the water for some hours we shall find it covered by a spongy ice structure. Ice which is growing on floating objects will be subjected to similar conditions, but not so extreme. In this way the running water will contain floating aggregates of growing ice sponges generally called frazil ice.

If the supercooling is sufficient and the turbulent action strong, there will also be formation of similar ice structures at the bottom of the river. In Norway and in northern countries generally the beginning of winter frost will very often be followed by the formation of this type of ice in the rivers. Under a hard frost the production may be very heavy and sometimes the river section may be partly filled with bottom ice, causing flooding of the river banks.

I have just emphasized the connection between the growth of the crystals and the movement of the supercooled water, and I mentioned the tendency which would exist for aggravating all peaks and sharp needles. This in its turn again will act as a brake upon the movement of the water and, keeping this in mind, we have the key to the understanding of the formation of the ice dams in running rivers. We may then observe the river running over an ice bed which has the appearance of a staircase. The front of each step is formed by an ice dam producing a water basin with a still surface. The ice dam has been built by the formation of bottom ice (anchor ice), the growth being most active where the velocity of the water is greatest. If the cold is persistent the dam may be built up in a comparatively short time.

By this means the turbulent river will be transformed into a series of dams, each with a comparatively still water surface which is very rapidly covered by a growing ice sheet. Provided that this process is not interrupted the loss of heat is slowed down and the formation of bottom ice will stop as soon as the water is covered by ice. Now two processes representing heat gain will more and more change the process. First, the radiation from sun and sky will balance part of the heat loss from the ice surface; and secondly, the water running underneath will be slightly heated due to the energy produced by its falling down the slope. Instead of supercooling we may even find a slight heating of the water raising the temperature above 0°C . and accordingly some ice will melt. This will take place where the water is running fastest and consequently the stream will cut its way through the ice bed and after some time run in a channel the bottom of which is the river bed. It will have ice walls on both sides and if the width is not too great

the ceiling will be formed by the ice cover of the previous basins, forming a sort of staircase.

This process is a remarkable example of how nature counteracts a disturbing process and re-establishes a new state of stability. If you are walking in the mountains, and I am sure you may observe it in England or Scotland during a hard winter you will find brooklets and small rivers running chuckling underneath a thick, cover of ice which will consist partly of a solid ice cover and partly of an ice structure which may sometimes remind you of a heap of rubble seen after an air raid.

In some rivers in both Norway and Sweden as well as in Russia and Siberia, the process which I have just described may be interrupted by a sudden change of temperature. If this takes place during the formation of ice dams, it may result in a breach of the dam. The water stored behind the dam will rapidly be released the next dam will in turn be emptied, and if the slope is sufficiently steep a flood wave will rush down the river sweeping with it the ice masses and even causing great damage. Where the river takes a more quiet course the ice masses will be stopped and may be pressed into barriers of considerable height. In more southern latitudes ice formations in rivers may lead to ice floods of a similar character, if not on such a scale. Where such a wave of water and ice, carrying masses of ice blocks, forces its way, overflowing the banks, you will see the eroding power in action, a nature's bulldozer on a gigantic scale, wherever an obstacle tends to stop the advance.

Some practical applications

The formation of frazil ice, anchor ice, and ice dams may very severely hamper both traffic on the rivers and the efficiency of water-power installations. The chief places of production of frazil ice and anchor ice will be the rapids. Here anchor ice will be formed, especially bellow the rapids, and the frazil ice will be carried downstream and deposited in huge masses underneath those parts which have been covered by a continuous ice cover. Compare Fig. 19, p. 114, in « Ice formation », by H. T. Barnes (Chapman and Hall, 1906).

The precautions which may be taken to keep traffic going will evidently have to follow the example of the nature itself : to reshape the river by reducing the rapids eventually concentrating them by damming the water at proper places, thus producing the staircase which I mentioned above. If traffic is to go on, evidently locks will have to be established if necessary.

Hydraulic power-plants in northern countries may often be exposed to difficulties caused by the formation of frazil ice and anchor ice. In many of the Scandinavian plants, this has led to the application of artificial heating of the grids which are placed before the inlet to the turbines. There is, of course, no attempt to bring the temperature of the supercooled water as a whole up to or above the freezing point. It will be sufficient to keep the surface of the grid slightly warmer than the water. It may however happen that this precaution will not be sufficient, and then it will be necessary to avert as much as possible the formation of frazil ice and bottom ice. From the analysis which we have given above of the nature of ice formation in running water, some basic principles governing the planning of precautions will clearly emerge.

First, the profile of any open channel leading water to the power plant must be sufficiently wide to reduce the turbulence of the water. Secondly, it might be advisable to accelerate the formation of an ice cover by placing wooden beams across the channel from which an ice layer could start growing upstream; this precaution is used in many cases. Thirdly, if the supercooling of the water cannot be fully avoided, it is of great importance to provide for a slow and smooth running of the water through the inlet channels to the turbines. Good results have been obtained by widening the inlet openings and shaping the walls and turnings as much as possible in a streamlined fashion.

In Norway I have also seen a few examples of bottom ice blocking the water supply delivered from a dam. Fig. 5 is a sketch of the reservoir supplying one part of Bergen in Norway. Some ten years ago when on a winter day a strong continental wind was blowing down the mountains and sweeping over the dam, the flow through the water main suddenly stopped. Nothing unusual could be seen on the dam, which was full of water with no ice on it. A diver went down to the inlet of the water pipe and found it completely covered by a thick crust of bottom ice. This was at a depth of 30 feet below the surface of the water. The production of this ice crust is easily understood : the water in the basin had been

cooled throughout to about freezing point, the cold wind had partly produced supercooling and simultaneously started circulation in the reservoir causing supercooled water to go down the back of the dam and return along the bottom. The grid shielding the inlet of the tube presented the suitable starting surface for ice growth with the result just described. In the climate of Bergen it is most improbable that such a thing should happen often, and accordingly the only precaution which was recommended was the placing of a floating chain of wooden beams to counteract the circulation and start the production of a shielding ice cover.

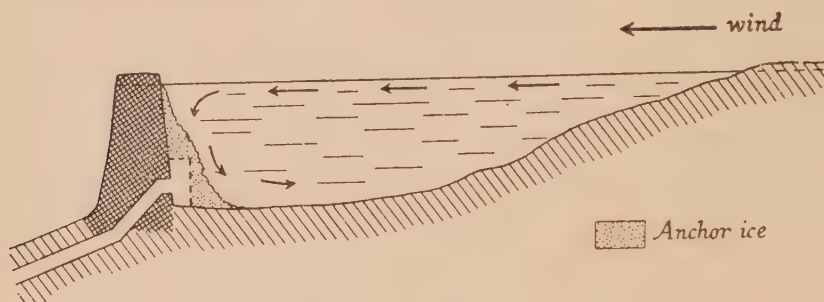


FIG. 5. Water main from reservoir choked by anchor ice.

The researches mentioned on p. 184 have been fully described in a paper printed in the *Geofysike Publikasjoner* in 1942. The study of the temperature of the surface layer directly in contact with the air is of interest not only in connection with problems of ice formation. It will enter into a detailed study of the exchange of heat between water and atmosphere, and the results of our measurements indicate that considerable corrections will have to be made in the measurements of water temperature with thermometers if we want to know the real temperature of the separating boundary layer.

2) LA GENESE DE LA GLACE FLOTTANTE ET SON APPARITION SUR LES COURS D'EAU DE L'EUROPE CENTRALE APPARTENANT AU BASSIN BALTIQUE

par

Ing. Dipl. **JULIAN LAMBOR**, Directeur des Voies Navigables à Poznań (Posen), Pologne

Le résumé du traité publié en 1948 No. 3 de l'édition périodique « Wiadomości Hydrograficzne ».

Parmi tous les phénomènes glaciaux sur les voies navigables intérieures, abstraction faite des engorgements des glaces, les hydrotechniciens relèvent le phénomène de la glace flottante (fraisil), surtout le moment de son apparition, car dès ce moment la morte saison commence sur les rivières. L'apparition soudaine du fraisil sur les cours d'eau rend impossible le retour des bateaux dans leur base pour leur toilette hivernale, en conséquence de quoi il y a de grandes pertes matérielles et de grands frais d'assurance, résultant du fait, que les bateaux restent en mi-route. C'est pourquoi on tâche de résoudre le problème, malheureusement avec peu de résultats jusqu'à présent, de la prévision de l'apparition du fraisil, cet avant-coureur de la congélation de nos rivières.

En mentionnant le fait de la prévision de l'apparition du fraisil — il faut se rendre compte du cours de ce phénomène et de sa genèse.

Il est facile de distinguer l'apparition du fraisil (en angl. : frazil, slush-ice, current-ice; en franç. : fraisil; en allem. : Treibeis; en russe : szuga, szoroh; en scand. : Issörja) tellement caractéristique pour nos rivières. Le fraisil se forme subitement sur la surface de l'eau courante, ressemblant à une masse spongieuse, composée de petits disques, avec une couronne blanche aux bords. Il apparaît sous forme de mottes irrégulières, composées de petites aiguilles de glace qui, en conséquence d'un frottement mutuel, forment de grands emplâtres avec une couronne caractéristique dont la couleur blanche provient de l'air qui remplit le vide de la partie au-dessus de la surface de l'eau. La partie moyenne est plate, imprégnée d'eau jaunâtre-bleue; quand la température baisse, elle gèle et se change en glace flottante; (en angl. : plate-ice; en allem. : Eisscholle).

Pour déterminer ce phénomène on a créé différents termes, parfois même contradictoires, car la formation du fraisil et sa nomenclature n'est pas encore claire. Surtout confond-on la conception de « fraisil » avec celle de la « glace de fond » (en angl. : ground-ice; en russe : gruntowoj led; en allem. : Grund-eis).



FIG. 1. Photographie du fraisil (photographié sur la Warta à Poznań, le 14-XII-46.

Nous avons plusieurs théories concernant la formation du fraïsil. Certains observateurs sont d'avis que le fraïsil au bord et sur la surface des eaux se forme par suite de l'action du vent, qui fait monter les parcelles de l'eau qui, se mêlent à l'air froid, forment des cristaux de glace. Les autres sont d'opinion que la glace flottante nait au fond de la rivière de la glace de fond, qui monte à la surface de l'eau, quelques-uns enfin, ne trouvant pas dans ces théories l'explication de différents doutes, supposent que le fraïsil se forme aussi dans le milieu de la masse de l'eau, mais trouvent, que sa formation est due à l'air froid qui, sous l'influence des courants turbulents pénètre dans les couches inférieures de l'eau et forme ensuite la glace dans toute la masse de l'eau. Les adhérents contemporains de cette théorie sont : Devik (Scand.) et Altberg (Russie).

Sans aucun doute la glace se forme sur la surface de l'eau et prend part dans la formation du fraïsil, mais pourtant pas dans de pareilles dimensions, qui pourraient expliquer ce phénomène en total. F. Reymann, p. ex., est d'avis, que la congélation a lieu seulement à la surface, mais cependant beaucoup de faits nient, que le fraïsil se forme uniquement à la surface.

La glace de fond participe dans la formation de la glace flottante. Un des précurseurs de la théorie de l'origine de la glace de fond fut Lord Kelvin et, après lui, un grand nombre de glaciologues de différents pays étudient ces théories. Aujourd'hui le fait de la formation de la glace de fond sur les rivières de l'Europe Centrale du bassin de la mer Baltique est suffisamment constatée — quoique on pourrait discuter le cours de ce phénomène.

Il est évident aussi, que le concours de la glace de fond dans la formation de la glace flottante est très grand, quoique pas exclusif.

La théorie de la genèse de la glace de fond, basée sur le rayonnement nocturne dans l'espace — de Kelvin, J. Farquharson et Barnes, ne fut pas niée (quoiqu'il y a aussi des opinions contraires, comme p. ex. celle de Arago) et elle a toujours beaucoup d'adhérents. — Par « rayonnement nocturne » on désigne le rayonnement d'un corps vers l'espace, moins le rayonnement que ce corps reçoit de l'atmosphère. Grâce à la science de physique nous savons que tout corps chauffé envoie des rayons caloriques quand sa température devient plus haute que celle de son milieu. Le rayonnement est d'autant plus fort, que la différence des températures est grande et en proportion de la quatrième puissance. — Plus la température d'un corps est élevée — plus les rayons sont courts. Les corps froids envoient des rayons invisibles (infrarouges) du rang des dixièmes d'un millimètre et des rayons plus longs (les rayons visibles sont de 0.0004 mm de longueur). Dans les heures nocturnes vient un grand débit de la chaleur, d'autant plus grand, que la différence des températures est grande et que l'atmosphère est plus pénétrable pour les grandes ondes du rayonnement calorique, à condition que le ciel soit clair et sans humidité. — C'est pourquoi le fraïsil se forme pendant les nuits claires et froides, ce qui est très caractéristique pour ce phénomène. Les ondes longues du rayonnement sombre passent par l'eau et d'autant plus facilement, que ces rayons sont longs, (Nichols, Aschkinass, Barnes). L'eau n'est pas un obstacle dans la formation de la glace de fond à cause du rayonnement du fond du fleuve. Les physiciens déterminent la grandeur de ce rayonnement de env. 0,17 cal/cm² par 1 minute.

Quoique l'énergie de ce rayonnement est beaucoup plus petite que celle des rayons courts — elle est suffisante pour abaisser la température de centièmes de degrés pour provoquer la cristallisation à la température de l'eau près de 0° (les centièmes et même les millièmes de degré) et non seulement au milieu de la masse de l'eau, mais aussi au fond du fleuve. Les mesurages des températures de l'eau dans différentes profondeurs ont permis de constater une même température dans le profil entier et les différences rares de 0,05° sont dans les limites du mesurage, contrairement à l'opinion présentée souvent, qu'au fond d'un cours d'eau la température est de 4°. Dernièrement des mesurages ont été faits sur le lac de Kiekrz et on constata une température égale jusqu'à la profondeur de 14 m. Ce phénomène est en rapport non seulement avec un mélange continu de l'eau — mais il est causé aussi par le rayonnement de la masse de l'eau. Ce mélange est encore plus intense dans une rivière.

Les phénomènes de glace diffèrent de ceux de l'eau, car les rayons caloriques ne peuvent y passer, surtout dans la glace de l'eau courante, qui a des qualités physiques semblables à la neige et souvent la glace est couverte d'une couche de neige. Avec le moment de la formation d'une couche de glace sur le cours d'eau — la glace de fond cesse de se former et celle qui s'est déjà formée monte à la surface.

L'opinion, que sur les lacs et les eaux stagnantes la glace de fond et le fraïsil ne

se forment pas, n'est pas juste ; ils s'y forment pourtant, mais sur une étendue plus petite et leur vue extérieure est différente — ils ne dépassent pas la surface de l'eau et deviennent une couche uniforme de glace. Nous observons la formation de fraïsil et de la glace de fond non seulement aux bords du Labrador, mais aussi dans le Kattegat et même dans le lac de Boden (Ellison Hawks : The book of air and water wonders).

Cette théorie de la formation de la glace de fond en conséquence du rayonnement nocturne est confirmée par des observations, donc elle est difficile à nier. En réalité la glace de fond naît pendant les nuits claires sous forme spongieuse et, pendant le jour, quand le soleil chauffe le fond, elle monte à la surface sous la forme de glace flottante. C'est donc une opinion juste de gens simples que la lune forme la glace de fond et le soleil la fait monter à la surface. Cependant la formation de la glace flottante de cette manière, ne résout pas entièrement ce problème et elle n'explique pas les circonstances suivantes :

Si la glace flottante se formait de cette façon il nous faudrait trouver dans le fraïsil — après son dégel — une grande quantité de sable et de bourbe. En effet il y a dans la glace flottante un peu de sable mais beaucoup moins, dix fois moins, que nous nous y étions attendus. Il est facile de distinguer le fraïsil né au fond, à simple vue, par sa couleur et sa structure ; il est aussi plongé plus bas dans l'eau. En effet nous observons une telle glace flottante en certaines compositions, mais pas plus grandes que 1/10 de toute la masse de la glace flottante. Nous ne la voyons presque jamais dans le fraïsil jeune, c. à. d. le premier jour de sa formation. D'où provient donc le reste de la glace flottante ?

En outre la glace de fond se forme plus facilement dans certains endroits du cours d'eau qui ont une spécifique disposition. En conséquence nous devrions observer, que dans certains endroits il s'en forme plus et plus intensivement, cependant la glace flottante étonne par sa régularité et par l'uniformité dans des grandes parties des cours d'eau ; cette régularité n'est jamais propre à la glace flottante née au fond. On observe distinctement la formation de la glace de fond au cours de l'action des brise-glaces sur les rivières du bassin de la mer Baltique.

De même le moment de la formation de la glace flottante n'est pas distinct. La glace de fond se forme dans la nuit, pendant le plus grand rayonnement terrestre et se détache du fond pendant la journée quand le soleil chauffe le plus. Elle devrait donc apparaître sur la surface à midi. Cependant la première glace flottante apparaît après le coucher du soleil, pendant la nuit, et le matin tout le cours d'eau en est couvert.

C'est pourquoi beaucoup d'observateurs trouvent que la glace flottante se forme en premier ordre dans toute la masse de l'eau. Dr. ing. Leppik accepte, après dr. Devik et Altberg, que la glace flottante se forme dans toute la masse de l'eau. En conséquence de la convection des courants turbulents — l'eau est en contact avec l'air froid et, grâce aux suspensions dans l'eau — elles facilitent la cristallisation de l'eau. Ces savants nient la théorie du rayonnement de l'eau et acceptent seulement la convection qui introduit l'air froid dans la masse de l'eau. Cette théorie cependant, est aussi trouble que les précédentes.

Si la glace flottante se formait de cette manière — le vent fort et turbulent et la température basse de l'eau devraient jouer un rôle important. L'heure du jour serait sans importance. Comment donc expliquer la première apparition de la glace flottante pendant la nuit et d'autant plus forte, que les conditions du rayonnement nocturne sont meilleures, c.à.d. le ciel clair et la différence des températures aussi pendant un calme absolu.

Si réellement la glace flottante se formait de cette façon, les meilleures conditions seraient sur les torrents, où le mélange de l'eau avec l'air froid est le plus fort. C'est là que la glace flottante apparaîtrait le plus intensivement et en premier lieu ; sur les eaux calmes elle apparaîtrait plus tard et plus faiblement — mais il n'en est pas ainsi.

Pendant l'écoulement de la glace flottante le mélange de l'air froid avec l'eau est le plus intense. Les masses spongieuses de fraïsil, remplies d'air glacial, se précipitent, plongent chaque fois dans l'eau, entraînent des particules d'air froid, — dans ce cas probablement le mélange des masses d'eau avec l'air froid est le plus intense et cela d'autant plus, qu'il y a plus de fraïsil. Nous en pouvons tirer la simple conclusion, que plus il y a de glace flottante — plus sa formation devrait être intense. Nous observons cependant le contraire : plus il y a de glace flottante sur la surface de l'eau — moins il s'y forme de nouvelle glace. L'intensité du fraïsil a des limites qu'elle ne franchit jamais. Nous n'observons pas de surcharge du cours d'eau par la glace flottante. L'intensité du fraïsil se règle elle-même malgré l'influence de différentes conditions atmosphériques. Ce phénomène

ne dépend que de l'action du rayonnement, qui diminue à mesure que la surface se couvre de glace flottante.

Après que le premier fraïsil a passé nous apercevons que le fraïsil suivant ne se forme pas facilement, contrairement à ce qu'on pourrait conclure d'après la théorie de convection. Si p. ex. un engorgement de glace se forme, l'eau pure coule au-dessous de celui-ci. Nous constatons alors que sur cet espace libre le nouveau fraïsil se forme avec difficulté, comme si le moment de la sensibilité était passé, malgré que les conditions thermiques et atmosphériques n'ont pas changé, ce qui peut être aussi expliqué par la théorie sus-mentionnée :

M'appuyant sur mes observations et mesurages je suis d'avis que la glace flottante se forme en effet non seulement à la surface ou au fond, mais surtout dans toute la masse de l'eau, ce qui avant tout produit l'intensité du fraïsil flottant à la surface du niveau, et que la cause de sa formation n'est pas le mélange de l'eau avec l'air froid qui a lieu, en effet, mais avec un bien faible résultat ; la cause de la formation de la glace flottante repose sur ce que les suspensions emportées dans l'eau perdent leur chaleur et se refroidissent intérieurement à la température de leur milieu grâce au rayonnement sombre. C'est donc le même processus que celui qui accompagne la formation de la glace de fond.

L'indubitable dépendance de l'intensité de la glace flottante du rayonnement terrestre d'une part et un relativement petit contenu de la glace de fond dans le jeune fraïsil d'autre part — confirment cette supposition : de même le difficile processus de la formation du fraïsil dans l'eau pure (lacs, torrents, fleuves de la plaine basse, immédiatement après le passage du jeune fraïsil) en donnent la preuve. En acceptant le principe généralement approuvé que la glace de fond se forme grâce au rayonnement froid, je ne vois aucun motif à supposer, que le même processus n'ait pas lieu dans les suspensions, que le cours d'eau mène en grandes quantités dans toute la masse.

L'influence des suspensions dans l'eau, relative à sa congélation n'est pas suffisamment constatée. Ceux qui examinent ces phénomènes attribuent le rôle excitant à un corps étranger, tombé dans un liquide refroidi : les autres trouvent que c'est le milieu de la cristallisation, quelques-uns supposent que les suspensions ne jouent pas ici le rôle immédiat, mais médiateur de condensateurs et que le processus de la congélation ne se produit entre elles. En effet la collaboration des suspensions facilite le processus de la congélation. Mais il est important, de quelle manière elle se produit.

Ainsi que tous les autres corps l'eau répand aussi la chaleur par un rayonnement égal de toute sa masse, car l'eau laisse passer facilement les grandes ondes du rayonnement nocturne. Chose comprise que les corps d'une couleur sombre, suspendus dans l'eau, répandront la chaleur plus intensivement que l'eau même, c'est pourquoi sur ou dans ces suspensions la congélation de l'eau courante aura lieu le plus tôt. Le courant d'eau, qui emmène la chaleur de la congélation, facilite énormément ce processus. Si nous disons que ces suspensions deviennent le milieu de la cristallisation de l'eau, — c'est ainsi qu'il faut interpréter l'action de ces milieux de cristallisation. Ce processus n'est donc pas identique avec le phénomène des excitants de la cristallisation cité plus haut.

Dès le moment de l'apparition des phénomènes glaciaux sur les cours d'eau — nous observons la purification systématique de l'eau courante. — C'est un fait constaté, que les suspensions qui forment le milieu de la cristallisation sont emmenées avec le fraïsil. Leur quantité dans l'eau courante diminue sensiblement, — cependant on se serait attendu à ce qu'elles se trouvent en grandes quantités précisément dans la glace flottante.

Des observations et des mesurages exacts furent faits sous mes auspices pour examiner les phénomènes mentionnés ci-dessus pendant l'hiver de 1946-47 sur le fleuve Warta. Par ces études a été observé le premier moment de l'apparition de la glace flottante, le moment de sa formation, ainsi que quelques semaines avant et après sa première apparition. Les études faites avaient le but de répondre aux questions relatives au début de la formation du fraïsil et des circonstances l'accompagnant et de la consistance physique de l'eau quant à la quantité et la qualité des suspensions, précédant l'apparition du fraïsil, comment ces conditions changent pendant la formation de la glace flottante et de l'étude de la structure et de la consistance du fraïsil. Outre l'étude des conditions atmosphériques dans toute la période on mesurait continuellement la température de l'eau, on faisait l'analyse de la quantité et de la qualité des suspensions et leur étude microscopique. Les résultats obtenus ont complètement confirmé ma théorie de la formation de la glace flottante dans toute la masse d'eau par le rayonnement obscur des suspensions, qui se trouvent dans l'eau.

Ces études furent faites sur la Warta (affluent droit de la Odra) près Poznań, où nous avons un bassin de 25.117 km², pente de 0.192 ‰ avec le débit de l'eau moyen annuel env. 100 m³/sec, à la hauteur de 2.0 à 3 m dans le profil étudié.

En résultat des mesurages faits plusieurs semaines avant l'apparition des phénomènes glaciaux on a trouvé les quantités moyennes suivantes de suspensions dans 1 m³ d'eau coulante, en dépendance de la hauteur de l'eau sur le limnimètre de Poznań :

Hauteur de l'eau	200	quantité de suspensions dans 1 m ³	→	100 cm ³
»	»	»	»	»
»	»	300	»	127
»	»	»	»	»
»	»	350	»	140
»	»	»	»	»
c.à.d qu'en moyenne — 120 cm ³ dans 1 m ³ .				



FIG. 2. Microphotographie de la matière trainée au fond du fleuve 100 fois agrandie.

Il faut mentionner, que le volume noté se rapporte aux suspensions dans l'état humide et pas serré. Après que ces suspensions avaient complètement séché, le volume diminuait plus de dix fois, la matière se ratatinait intensivement. Le poids spécifique de ces suspensions détachées indiquait 0,065, et dans l'état sec il dépassait peu l'unité. En examinant ces suspensions dans le microscope nous voyons un tout autre tableau, que celui de la matière trainée.

Il suffit de comparer la fig. 2 et 3.

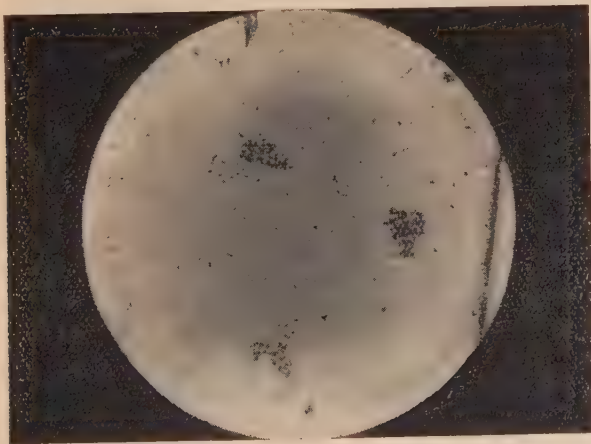


FIG. 3. Microphotographie des suspensions du fleuve Warta, 350 fois agrandie.

Comme il résulte des mesurages et des calculs nous trouvons dans 1 cm³ d'eau env. 6×10^{-5} cm³ de suspensions, d'un poids total de $7,8 \times 10^{-6}$ g. Leur grandeur, établie dans le microscope, est dans les limites de 0,0002 - 0,003 mm. Nous trouvons des grandeurs de 0,01-0,02 mm en 10 % et seulement exceptionnellement encore plus grandes. Le diamètre moyen, résulté des mesurages proportionnellement à la masse de chaque parcelle est de 0,025, le volume de telles parcelles, est de $1,2 \times 10^{-7}$ cm³. Dans chaque 1 cm³ de l'eau il y a en 0,06 mm³, c.à.d. 3800 pièces, d'après leur grandeur moyenne déjà acceptée. Imaginons qu'elles sont réparties régulièrement dans la masse de l'eau — elles seront disposées alors à une distance de 0,64 mm et chaque parcelle serait entourée d'une masse d'eau du volume de 0,264 mm³.

Ces suspensions de couleur sombre, emportées dans l'eau courante, envoient dans l'espace des rayons caloriques d'une grande onde; l'eau ne présente pas d'obstacles et laisse passer les rayons caloriques d'autant plus facilement que leurs ondes sont plus grandes. La grandeur de ce rayonnement, d'après les mesurages de Angström et Haurer, peut être établie à 0,11 cal/cm²/min., ayant pris en égard la chaleur venue de l'atmosphère dans la nuit. Si la surface de ces corps est en moyenne de $6,2 \times 10^{-6}$ cm², nous obtenons la valeur du rayonnement dans 1 heure, 4×10^{-5} cal/heure. Ce n'est pas peu, car la température de l'eau — si elle est quelques centièmes ou millièmes de degré au-dessous de zéro — cause le changement de la masse de l'eau de volume 5×10^{-7} cm³ en glace en une heure. Ce petit glaçon devient le centre de la cristallisation. Dans chaque mètre cube d'eau courante peut se former de cette façon 1 dm³ de glace dans une demi-heure, qui monte ensuite à la surface sous forme de fraïsil. C'est de là que vient la rapide apparition d'une grande masse de fraïsil.

Rien ne nie le procès mentionné plus haut, — au contraire — tous les phénomènes le confirment.

La quantité de glace flottante ainsi produite, se règle d'elle-même. A mesure que cette quantité augmente, le processus de sa formation s'arrête, car la glace flottante, de même que la neige ne laisse passer les grands rayons et s'affaiblit avec le manque de suspensions.

Les mesurages faits sur la quantité des suspensions dans le fraïsil, où elle devrait être beaucoup plus grande que dans l'eau courante, ainsi que ceux faits sur le manque de suspensions dans l'eau — ont entièrement confirmé cette supposition. — Nous avons trouvé en effet une grande condensation des suspensions dans le fraïsil et c'est justement de celles, qui étaient observées d'avance dans l'eau courante. Des échantillons pris dans le moment de la première formation du fraïsil le soir du 13-XII-1946 ont donné :

petit sable de fond	253 cm ³
solides au-dessus de 0,4 mm	487 g
suspensions	564 cm ³

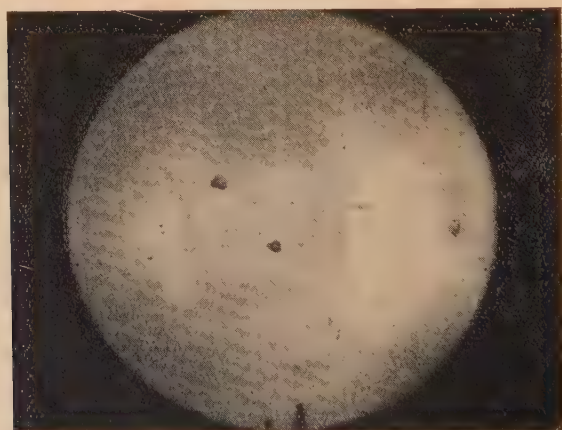


FIG. 4. Microphotographie de l'eau de la Warta (vue normale) avant l'apparition de la glace flottante (agrandie 60 fois).

Ce qui attire notre attention, c'est comparativement une petite quantité de sable de fond, notamment seulement 253 cm^3 conformément à l'affirmation, que la glace de fond ne prend que peu part dans la formation de la glace flottante et elle est postérieure, car elle apparaît à la surface seulement le jour suivant, où sa quantité augmente jusqu'à 1420 cm^3 et dans quelques jours elle diminue et oscille dans les limites de 820 cm^3 à 960 cm^3 , ce qu'il faut attribuer à la difficulté de la formation de glace de fond, causée par la glace flottante qui couvre la nappe d'eau.

Les solides au-dessus de $0,4 \text{ mm}$ de diamètre sont des parties du monde végétal, qui flottent sur la surface de l'eau et s'attachent en gèlant à la glace flottante.

Les petites suspensions, qui avaient été observées avant dans l'eau courante, furent trouvées dans le fraïsil jeune en quantité de 564 cm^3 ; le 3^e jour, quand la marche de la glace flottante était beaucoup plus intense, cette quantité augmente jusqu'à 1080 cm^3 , c.à.d. sa condensation fut dix fois plus grande que dans l'eau.

La vue microscopique de l'eau, obtenue du fraïsil en dégelé en comparaison avec la photographie de l'eau avant l'apparition du fraïsil — montre la condensation visible des suspensions dans le fraïsil. (voir fig 4 et 5).

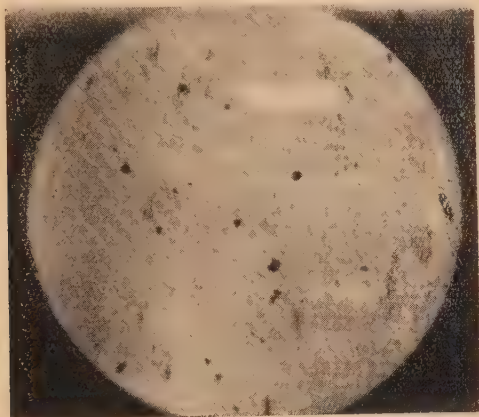


FIG. 5. Microphotographie de l'eau du jeune fraïsil (agrandie 60 fois).

Le centre de la cristallisation est trouvé très rarement pendant les études microscopiques (voir fig. 6), ce que prouve que ce sont les suspensions qui jouent ce rôle dont la vue caractéristique est présentée sur la fig. 3.

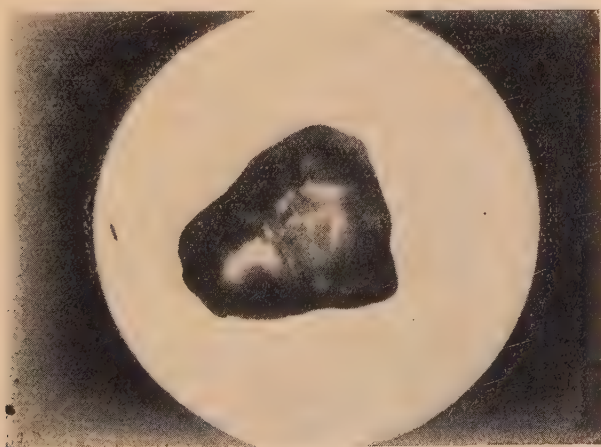


FIG. 6. Centre de la cristallisation, agrandie 100 fois.

Il faut ajouter que le débit de la glace flottante, même avec une petite quantité de glace de fond, n'influence nullement l'épuration de l'eau — juste le contraire.

On pouvait facilement voir dans la glace flottante les condensations de la glace de fond, qui se distinguaient par leur vue extérieure et par leur consistance et formation. On peut définir cette quantité à 10-20 % de la masse de la glace flottante. Il est caractéristique qu'on a observé le plus de la glace flottante le 3^e jour, cependant le 1^{er} jour, surtout au moment de l'apparition du fraïsil, on n'a pu voir de glace de fond. Les échantillons pris de pareilles condensations de la glace de fond, montraient le contenu du sable et de tourbe de 20 à 30 litres dans un 1 m³. Cette quantité n'est pas trop grande, prenant en considération que 1 m³ de glace peut emporter 90 kg.

Pour bien s'orienter dans le cours des phénomènes de l'apparition de la glace flottante dans le profil observé du fleuve Warta près de Poznań — il faut encore mentionner les conditions dans lesquelles l'apparition avait eu lieu. La première apparition du fraïsil a été observée le 13-XII-1946 le soir, quand l'eau avait la température de 0° (mesurée avec exactitude jusqu'à 1/10°) et la température de l'air était de — 8°C. Le 13-XII à 8 h. du matin la température de l'eau baisse subitement jusqu'à 2°C. et la température de l'air à 7°C.. A 13 h. la température de l'eau baisse jusqu'à 0°. Il est évident qu'avec cette température de l'air et de l'eau et par un temps clair — le moment de l'apparition de la glace flottante commençait ce qui fut signalé dans le cours supérieur de fleuve à la distance de 170 km. A partir de ce moment l'eau était soumise à une stricte observation pour pouvoir attraper le premier moment de l'apparition du fraïsil. Dans l'après-midi de ce jour on n'avait constaté aucun changement sur le fleuve. A 16^h l'eau commençait à s'épaissir sur la surface et à 16^h30 — le fraïsil se forma et avait occupé dans une demi-heure 1/5 de la surface du fleuve. Le lendemain matin la glace flottante occupait 2/3 de la surface et formait de grands emplâtres. La congélation du fleuve avait eu lieu pendant la nuit du 18/19.XII avec la température de l'air — 11° et par un ciel sans nuages, ce qui avait causé une grande condensation du fraïsil et la congélation complète du fleuve.

Des échantillons pris de la masse de fraïsil ne montrent en aucune façon la structure spongieuse ni ne ressemblent à du gruau, ce que l'on dit d'ordinaire, mais ils se composent de paillettes séparées. Ces paillettes cassées, rangées en toutes directions et mêlées donnent l'effet d'une masse spongieuse. Déjà Westman décrivait la structure pailletée du fraïsil et il prétendait, que ces paillettes changent ensuite en de courts et de gros pistons, grains et aiguilles, ce que personnellement je n'ai pas constaté. On avait essayé de saisir le fraïsil à la profondeur du fleuve au moment quand il naît, mais on n'avait pas réussi, malgré l'emploi des appareils spéciaux, à constater l'origine de cette glace. Il faut croire que la glace flottante est tellement fine, avant de monter à la surface, qu'il est difficile de l'observer.

En résumant les observations ci-dessus — il faut constater ce que suit :

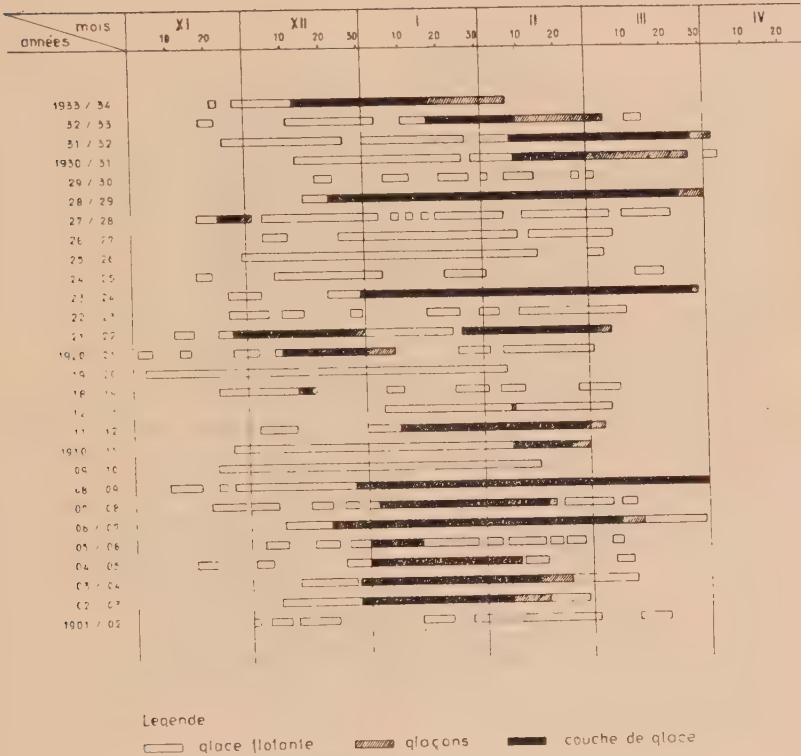
La glace flottante se forme dans toute la masse de l'eau courante en conséquence du rayonnement des suspensions que contient l'eau et de l'eau elle-même et ça c'est la cause principale de la formation de la glace flottante. En même temps, grâce au rayonnement nocturne, il se forme dans les lieux surtout disposés, la glace de fond qui, en montant à la surface de l'eau, flotte aussi en forme de glace flottante, mais en quantités beaucoup plus petites. La glace flottante, se montre à la surface de l'eau après la formation du fraïsil, né de la masse de l'eau. La participation de la glace flottante qui se forme de la masse de l'eau par le refroidissement de l'eau, causé par la convection de l'air froid, est insignifiante. La plus grande intensité de la formation du fraïsil est pendant le premier jour de son apparition, à mesure que le cours d'eau se couvre de glace de fond, — la formation du fraïsil cesse. Les conditions de la formation de fraïsil sont : La température de l'eau au-dessous de + 0.05°, la temp. de l'air au-dessous de — 5° et l'atmosphère pure. En rapport avec le sus-dit il faudrait établir la nomenclature : « glace flottante » se rapporte à toute glace qui flotte sur les cours d'eau en consistance libre, celle qui naît au fond et celle qui se forme dans la masse de l'eau. Le nom « fraïsil » est donné à la glace flottante qui vient de la masse de l'eau — pour la distinguer de la glace de fond.

La glace de fond et le fraïsil apparaissent sur tous les cours d'eau du bassin balteique, pas seulement — comme on le dit souvent, dans les pays situés tout au nord. Parmi tous les cours d'eau de l'Europe le fleuve Vistule a la période la plus longue de la marche de la glace flottante. Sur les cours d'eau de l'Europe Orientale le fraïsil se forme plus tôt, mais l'eau se couvre très vite d'une couche de glace. La quantité de jours dans une année, de la marche de la glace flottante sur la Vistule est en

moyenne de 56 jours, sur le Niemen et la Pregola — de 18 jours et sur la Odra et Elbe — 20 jours. La Vistule est le royaume de la glace flottante et c'est là que l'on voit le mieux ce phénomène et l'on ressent ses résultats néfastes pour la navigation.

La prognose des phénomènes glaciaux serait d'une grande utilité mais ce problème n'est pas facile à résoudre et en réalité on ne prévoit que les conditions atmosphériques. Jusqu'à présent ces efforts n'ont pas donné de résultats positifs. L'hydrologue russe O. Maszkiewicz (1936) présente une série de rapports, concernant les données statistiques de ce phénomène — mais sans dénouement final.

Phénomènes glaciaux dans les particulières années
Fleuve Wista — limnimètre Toruń

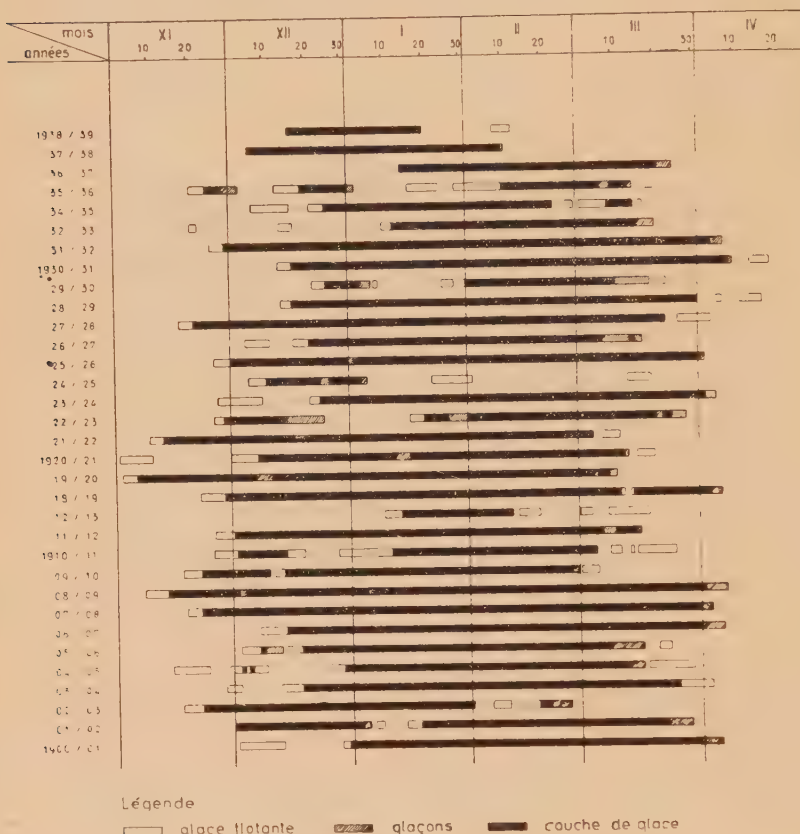


En comparant les tableaux de l'apparition des phénomènes glaciaux sur tous les cours d'eau de l'Europe — nous voyons une certaine conformité du cours de l'apparition de la glace flottante de l'est à l'ouest, produite par des conditions atmosphériques normales, qui provoquent l'apparition de la glace flottante. — Ainsi, p. ex. l'Elbe près de Darchau en rapport au Niemen près de Tilsit montre 4 semaines de retard en moyenne. L'apparition de la glace flottante sur la Vistule près de Toruń est un avertissement pour toute la Vistule et pour la Odra, car elle y apparaît 1-2 jours plus tôt que sur la Vistule près de Varsovie et 5 jours plus tôt que sur la Vistule près de Cracovie, quoique ces endroits sont situés plus haut et ont des conditions atmosphériques pires.

Je suis d'avis que seulement d'après ces observations la prognose de l'apparition de la glace flottante peut donner de bons résultats. Cependant pour une telle élaboration il faudrait avoir des matériaux statistiques qui malheureusement manquent dans les publications européennes hydrographiques. Les dates de la première

apparition de la glace flottante ne sont pas précises dans les annuaires hydrographiques de tous les pays. En outre, différents phénomènes sont marqués dans les annuaires d'un même signe; p. ex. le fraïsil, la glace de fond, la glace de bord, la glace flottante et le glaçon ont le même symbole dans les annuaires allemands et, par conséquent, ces observations sont inutiles. Le moment tellement important de l'apparition du fraïsil et de la température de l'eau n'est pas cité avec précision dans les annuaires de différents pays.

Phénomènes glaciaux dans les particulières années Fleuve Niemen - limnimètre Tylza



Ce problème peut être résolu seulement avec la collaboration internationale. L'Association Internationale de l'Hydrologie Scientifique est la plus propice pour établir des bases de collaboration uniforme.

Ci-dessous je propose les postulats au minimum qu'il faudrait pour présenter aux auteurs de publications des phénomènes glaciaux, comprises dans les éditions du service hydrographique des pays Baltiques, pour que ces publications puissent servir de matériaux statistiques, afin de dénouer la question des phénomènes glaciaux. Voici ces postulats :

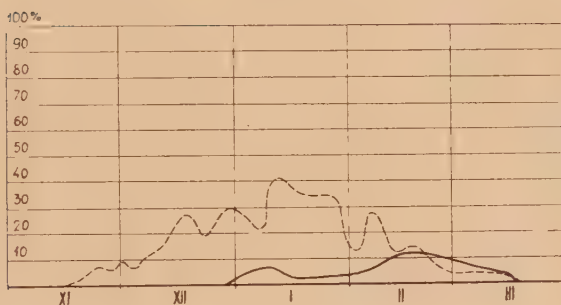
1. Noter par jour la température de l'eau dans plusieurs emplacements du bassin de chaque fleuve, en comptant un emplacement sur 300 km du cours du fleuve.
2. Noter par jour la température de l'eau avec exactitude jusqu'à 0.1°C en principe, mais la température au-dessous de 2°C doit être notée avec exactitude jusqu'à 0.01°C .

LA DURÉE DES PHÉNOMÈNES DE GLACE PENDANT LA PÉRIODE HIVERNALE /XI-III/ EN POUR-CENT DE TEMPS

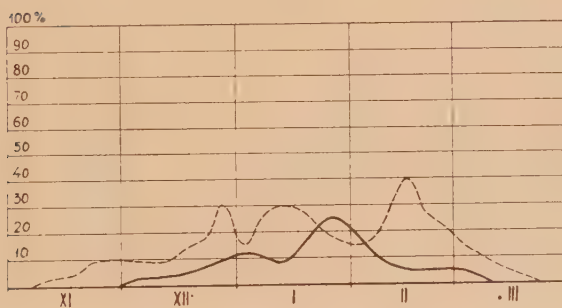
/Moyennes des années 1900-1940/

— Couverture de glace
----- Glace flottante

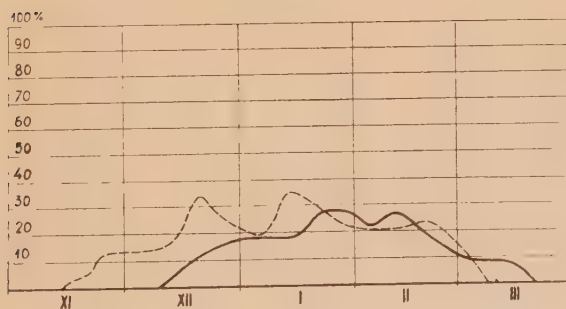
TARGAU ŁABA



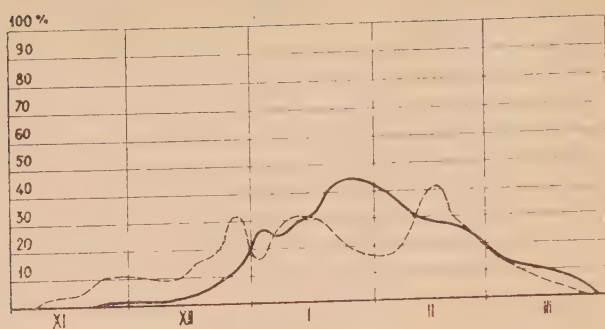
DARCHAU ŁABA



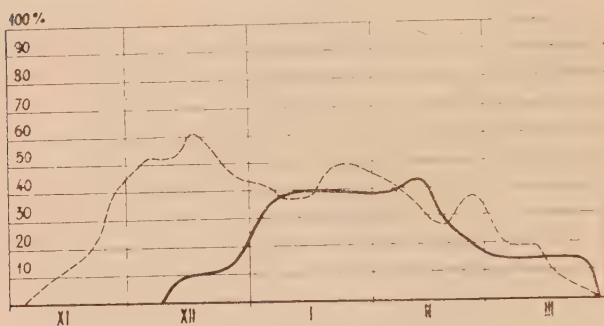
SŁUBICE ODRA



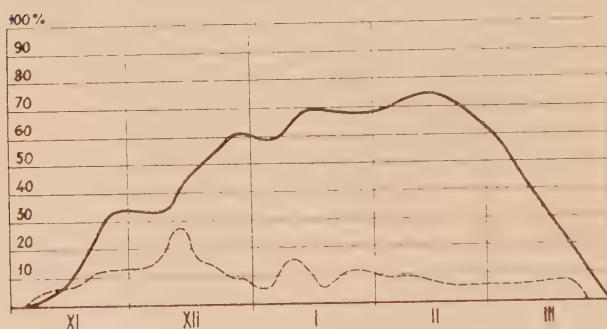
GORZÓW WARTA



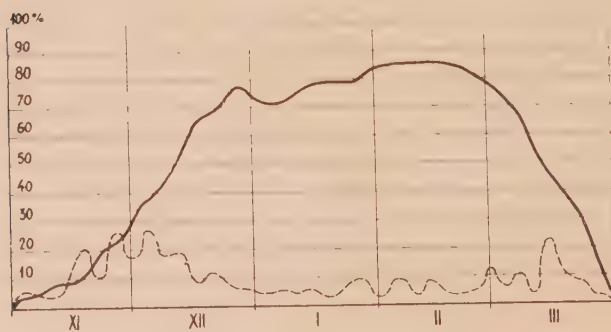
TORUŃ WISŁA



ŁAWA PREGOLA



TYLZA NIEMEN



3. Marquer la glace flottante par un signe spécial, différent de celui du glaçon qui s'écoule après la débâcle.

4. Indiquer exactement la date du commencement des phénomènes glaciaux avec une grande exactitude à propos du temps du jour sinon l'heure.

5. Ne pas marquer la glace de fond fixée au fond par le même signe que la glace de fond qui flotte détachée du fond, ou la glace flottante, surtout pas par le signe accepté pour la glace de bord.

6. Introduire dans la nomenclature des phénomènes glaciaux et des signes acceptés un système uniforme de définition de ces phénomènes.

M'appuyant sur le système de signes déjà accepté je propose les signes suivants:

- a) glace flottante, fraisil, glace de fond flottante;
- b) débâcle, glaçon (allemand. Eisgang, Eisscholle, Eisbewegung) en conséquence du mouvement de la couche de glace;
- c) glace de bord et glace de fond fixée au fond (anglais. ground-ice, anchor-ice), pas flottante;
- d) le même que c) mais au cas que ces phénomènes ont l'influence précise sur le débit de l'eau dans le fleuve;
- e) partielle couche de glace;
- f) congélation totale, couche de glace (allemand. Eisstand). Il faudrait bien par l'épaisseur de la ligne indiquer l'épaisseur de la couche de glace en trois nuances: moins de 10 cm, 11-20 cm, plus de 20 cm;
- g) engorgement de glace (allemand. Eisstopfung, Eisverstetzung).

7. Les signes qui définissent les phénomènes glaciaux doivent être marqués à côté des observations limnimétriques quotidiennes et aussi de la température de l'eau, ce qui a lieu dans les annuaires allemands.

8. Il faudrait indiquer la durée de la glace flottante (nombre de jours de l'année en somme), la durée de la couche de glace et en particulier la durée de la débâcle à partir du premier mouvement de la glace.

9. Un résumé des phénomènes glaciaux sur tous les fleuves et dans tous les emplacements, fait en aperçu sur une feuille de papier, ce qui a lieu dans les annuaires hydrographiques polonais, est bien utile et serait à souhaiter.

Résolutions

1. L'Association Internationale de l'Hydrologie Scientifique reconnaît la nécessité d'attirer l'attention des savants hydrologues sur l'importance de l'élaboration de la prognose des phénomènes glaciaux sur les cours d'eaux de l'Europe Centrale du bassin Baltique.

2. L'Association Internationale de l'Hydrologie Scientifique reconnaît qu'il est bien à souhait d'indiquer dans les publications hydrographiques la température de l'eau par jour, du moins pour plusieurs emplacements du bassin d'un fleuve. Le mesurage de la température de l'eau par la chaleur de l'eau au-dessous de 2° C doit être rendu exact jusqu'à 0.01° C. En outre, il est nécessaire d'augmenter l'exactitude de l'indication des phénomènes glaciaux en ce qui concerne le temps de l'apparition ainsi que la sorte du phénomène, en introduisant des symboles uniformes. De même il faut indiquer le commencement d'un phénomène avec exactitude jusqu'à une heure.

Ing. Lambor

Poznań, le 6 Janvier 1947.

SUPERCOOLING AND ICE FORMATION IN OPEN WATERS

by

OLAF DEVIK

Summary

The supercooling of water plays a transient, but important part in static ice formation in still water, in so far as it necessarily occurs when the freezing process of the surface layer is being started. It is of lasting importance, however, when the process of dynamic ice formation takes place in running water. The present paper concerns measurements of the supercooling in both cases, the chief available material concerning static ice formation, comprising records of the surface temperature by means of the outgoing radiation, observed by a Moll thermopile combined with photographic registering. The instrumental equipment is described, and examples are represented of surface temperature diagrams related to static ice formation. These records so far represent a documentation of a supercooling of the surface layer of open water descending below -1°C . It is emphasized that the state of supercooling is a stable one, the formation of ice crystals being dependent upon the existence of nuclei in the water, or of solid boundary surfaces from which crystallization will start and the liberated heat of crystallization will flow. The process of thermal conduction necessarily needs time, and thus a state of supercooling may be retained outside the crystal boundary surface, which passes more or less slowly into the adjacent medium as the freezing process goes on.

1. In a special treatise (1) concerning the thermal and dynamic conditions of ice formation in rivers, the author has considered the ice formation both (a) in still or slowly moving water, which was termed *static ice formation*, and (b) that in running, turbulent water, which was termed *dynamic ice formation*.

In the case of static ice formation a surface cover is formed, the crystallization thereby starting partly from solid matter at the beach, and partly from solid material suspended or floating in the water, especially from the ever-occurring suspended particles which serve as nuclei. If the weather is calm and the water at rest, everyone who has seen a pool being covered by ice will have perceived the network which is formed by crystals beams, leaves, feathers, stars and other forms, which extending from the starting places, leave the water open for a time between the elements of ice structure. It will be shown in this paper that the ice formation in this case is regularly preceded by a supercooling of the uppermost water strata, which will continue to exist at the open parts of the water until they are all covered by ice. The transmission of heat from the water underneath, to the surface exposed to loss of heat into the surrounding atmosphere, will be chiefly due to conduction, the water particles of the surface layers approximately keeping their places. The thermal conductivity for water at rest being small, a considerable temperature gradient may accordingly be established in the water layer immediately below the surface of the water. Thus temperatures of the surface itself extending to below -1°C may be observed (section 3).

In dynamic ice formation in running water, quite different conditions prevail. As long as the water is open, the individual particles of the water surface layer exposed to heat loss will be interchanged with others, within a period of time which is all the shorter the more turbulent the motion. The heat transmission in this case will be chiefly due to convection and mixing, and the thermal conductivity will greatly exceed that for resting water. Consequently, the temperature gradient maintaining the heat transmission from the deeper strata will generally be small, and observations at different depths will show small differences of temperature.

When the whole mass of water has been cooled to nearly 0°C , the ice formation begins, but a coherent ice cover is generally not formed, the typical process in northern latitudes being the formation of ice particles distributed in the whole water

(1) Olaf Devik : Thermische und dynamische Bedingungen der Eisbildung in Wasserläufen, auf norwegische Verhältnisse angewandt, Geof. Publ. IX, No. 1, 1931.

flow, loose drifting aggregations of such particles and often the formation of anchor ice at the bottom of the river. In the paper mentioned (p. 82-92) the author has treated the ice formation in this case, which is dependent upon a supercooling of the whole water mass. *Altberg* was the first to emphasize the importance of supercooling in the formation of anchor ice and has reported measurements confirming this. As an example, we may quote a paper by *Altberg* (1) reporting measurements of the *Neva*. *Altberg* then observed that the river water during the ice formation process was supercooled both at the surface and at the bottom, as well as in the intermediate layers. The supercooling was usually small, and would generally vary between 0.001°C and 0.01°C , rising in exceptional cases to as much as 0.1°C . During the period of the observations, there commonly occurred formations of loose ice under water, on submerged objects and sometimes at the bottom. In section 5 below we will consider temperature measurements in dynamic ice formation more closely.

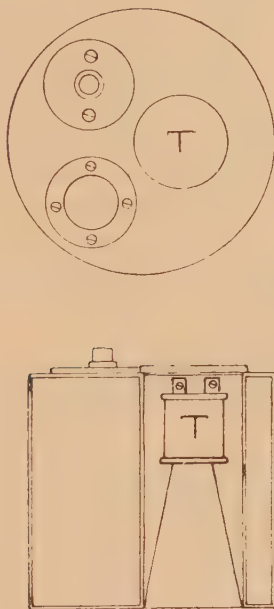


FIG. 1. Mounting used 1930. T-Thermopile.

2. A preliminary measurement of the supercooling occurring in open water in static ice formation was performed by the author in 1930 by means of the heat radiated from the surface. As receiver was used a Moll thermopile (Fig. 1), which was mounted in a doublewalled vessel, the space between the walls being filled with ice and water. The mean temperature of the thermopile was accordingly considered to be approximately zero. The potential of the thermopile was measured by a simple bridge coupling.

Several years later an error concerning the determination of the thermal sensitivity of the receiver was discovered, the numerical values of the surface temperature calculated being wrong and a critical revision of the arrangement of the thermopile made a new investigation desirable. The chief points concerning the revised apparatus will be mentioned here.

The receiving instrument of Fig. 1 is placed in cold air just over a much warmer water surface and strong temperature gradients thus characterise the immediate surroundings of the instrument. Moreover, the temperature distribution will change with every breath of wind, however small it may be. The heat transmission from the receiver instrument to the surrounding air will be considerable in the case

(1) W. J. Altberg: Anchor Ice. On the Cause of the Formation of Ice at the Bottom of Rivers and Lakes, Quaterly J. of the Royal Met. Soc. Vol. XLIX, No. 205, 1923

of Fig. 1, the receiver being approximately kept at the same temperature as the water surface. Of special influence to the measurements will be the heat current passing from the wall of the cone to the air.

The cone of the thermopile having a thickness of only one millimeter, considerable temperature waves may be propagated to the inner parts of the thermopile where the two groups of thermocouples (80 elements of constantan-manganin) are placed. To form an estimate of the influence of disturbing heat currents entering the material surrounding the two groups, as compared with the influence of the radiation changes which are to be measured, we first notice that the E. M. F. of the Moll thermopile is about $10 \cdot 10^{-3}$ volts when the difference of temperature between the two groups of elements is 1°C . Next we have to consider, according to section 3 below, that an E. M. F. of $1.90 \cdot 10^{-5}$ volts will be generated by the thermopile when it is exposed to the radiation from an extended water surface with a temperature differing 1°C from that of the thermopile. A temperature difference between the two groups of elements of the thermopile will thus produce an E. M. F. which will be about 150 times greater than that produced by an equal temperature change of the radiating surface. Consequently, if we wish to measure the surface temperature of water by radiation with an accuracy of say 0.05°C we shall have to keep stray heat currents in the material surrounding the thermocouples, so small that the temperature difference caused between the said groups keeps below 0.0003°C . It is evident that this will be difficult to attain even if the mounting of the thermopile be constructed in a way allowing only slow and small temperature changes to take place during the observation period. To obtain reliable measurements we shall also have to make such arrangements that a correction, if necessary, may afterwards be possible.

With a view to these claims, we abandon the idea of keeping the receiver instrument at a temperature differing much from that of the air (as in the preliminary arrangement of Fig. 1), letting the receiver with its mounting obtain the mean temperature of the surrounding air.

Summarizing we conveniently provide for :

- a) considerable heat capacity of the receiving system as a whole,
- b) good heat conduction in the material of the cone, in order to reduce temperature gradients that might sensibly alter the temperature distribution of the inner parts,
- c) effective shielding of the thermopile itself,
- d) diaphragms placed in the receiving cone to smoothe rapid air movements in the cone, constructed with due regard to b),

In order to be able to eliminate a moderate drifting of the galvanometer reading,

- e) photographic registering
- should be provided for.

3. A mounting of the Moll thermopile, which has proved to be satisfactory for observations in open air, is given in Fig. 2. The copper cone with a series of diaphragms is machined from a solid copper tube with thick walls. The connection with the inner tubing is designed as a thread coupling, with spare room for a window of rock salt or fluorspar. In calm weather or wind velocities below 1-2 Beaufort, however, no window is necessary when photographic registering is used. Almost all the measurements given below are made without a window. The thermopile itself is enclosed in a cylindrical copper chamber with thick walls, closed by an ebony cover, through which the conducting wires are passed. As a mantle there are used two copper cylinders, in order to screen away undesirable radiation and reduce the influence of the temperature gradient of the air outside the mantle. The cylinders also serve as ventilation tubes, being connected with a small electrically driven ventilator at the top of the cylinders. Ventilation is used to accelerate the heat exchange when the temperature of the receiver differs too much from that of the surrounding air; the ventilator is stopped, however, when observation begins (1). By this arrangement, the temperature of the thermopile receiver will approach the average temperature of the air column which it intersects at some distance from the water surface.

(1) To control the temperature of the inner parts of the thermopile, two special thermocouples are mounted, at convenient places. Before measurement begins it may then be ascertained, according to experience, whether disturbing temperature gradients are sufficiently low.

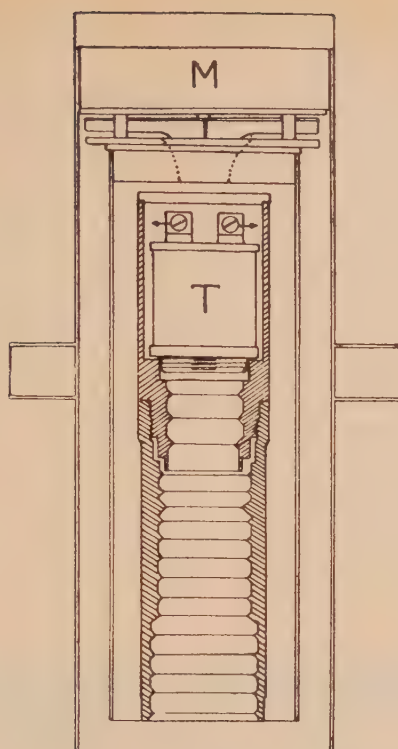


FIG. 2. T-Thermopile. M Motorhousing for ventilator.

When the receiver is exposed to the radiation from the water surface, the thermopile will generate an E. M. F. = V , proportional to the total temperature difference T between the water surface and the thermopile. We are, however, interested in registering only comparatively small changes of temperature of the water surface, with maximum sensitivity of the arrangement, and accordingly it will be necessary to compensate for the E. M. F. corresponding to the mean total temperature difference without diminishing the sensitivity of the galvanometer. Popularly expressed, we are to force the deviated galvanometer light spot back to the starting point,

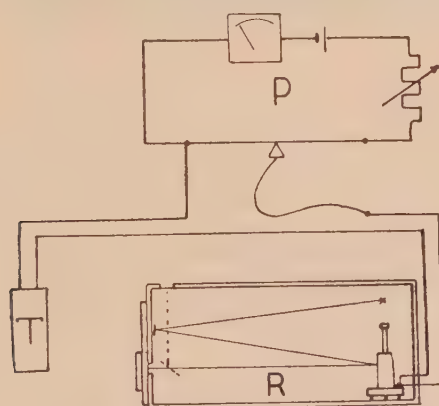


FIG. 3. T Thermopile. P potentiometer. R Registering case.

either by torsion or most conveniently by introducing a compensating E. M. F. opposite and equal to V , as indicated by the potentiometer circuit given in Fig. 3. The potentiometer reading of the potential V will then be a measure of the mean temperature difference T .

When the receiver is at temperature equilibrium with the air, T will also be equal to the difference between the temperature of the water surface and that of the air surrounding the receiver, which may be observed directly with a thermometer. Doing this we obtain a control of the calibration of the instrument. An independent calibration is given below.

A Moll galvanometer was used as recording instrument, mounted in a transportable case containing light beam arrangement, a transparent scale for visual observation and photographic registering arrangement (Fig. 3). A plate registrator was used, driven by a gramophone motor or by a synchronous electric motor. For time marking, the light beam is reflected aside to the transparent scale for visual observation, thus giving simultaneous control observation.

The calibration of the receiving apparatus was performed by means of water flowing rapidly over a plate placed horizontally underneath the receiver, at the same distance as that generally used for the actual measurements. By using alternatively water from two reservoirs of different temperature the simultaneous registering of the amplitude of the galvanometer may directly be turned into a temperature scale. According to calibration, the scale value on the original registering curve is $ds/dT = 18.1 \text{ mms}/^\circ\text{C}$, denoting by ds the scale difference of the galvanometer reading and by dT the corresponding temperature change of the radiating water surface. The practical limit of reading being $\frac{1}{3}$ mm, it will thus be possible to determine the temperature of the radiating water surface with a maximum accuracy of 0.02°C . On the registering curves reproduced in Fig. 5 below, the temperature scale is added,



FIG. 4. Thermopile receiver mounted at Sørnesset 1941.

From the scale value thus determined is calculated the sensitivity $dV/dT = 1.90 \cdot 10^{-5}$ volts per degree centigrade, representing the change dV of E. M. F. generated by the thermopile for a temperature change dT of the radiating water surface (1).

The calibration thus attained applies, as a matter of course, to temperature measurements concerning a water surface. Within the actual accuracy of the reading it will, however, also be applicable to temperature measurements of an ice surface: neither was an appreciable change of the reading observed when water, cautiously kept at 0°C , was allowed to freeze, and was covered by a thin membrane of ice, nor when the ice cover was perforated allowing water to spread rapidly over it.

(1) We have $\frac{dV}{dT} = \frac{dV}{ds} \frac{ds}{dT} = \frac{R_g + R_t}{R_g} \frac{dV_g}{ds} \frac{ds}{dT}$, R_g being the resistance of the galvanometer, R_t that of the thermopile, dV_g/ds the known volt-sensitivity of the galvanometer, and ds/dT known by calibration.

4. During April-May 1941 a series of observations were made at *Sörnesset* near *Atnasjøen*, at a height of 750 m above sea level. The ice formation was studied in a wooden water basin ($2\text{ m} \times 1\text{ m} \times 0.5\text{ m}$) through which water from a spring was slowly passing. Both at the inlet and at the outlet of the basin the water remained open, while the central part received a covering of ice. Here the receiver was placed 10 cm's above the surface (Fig. 4), the observations being made in the evening or in the early morning hours, when the temperature was at its lowest. At this time of the year temperatures below zero at this place occurred with clear sky or light clouds.

At every registering it is necessary to have at least one observation of the radiation from a 0° surface at the place of the water surface. The 0° surface is procured either by a freshly prepared snow stirabout, or more conveniently by the thin membrane of newly formed ice, the temperature of the top surface of a leaflet of floating ice being very nearly 0°C as will be shown below.

It is to be observed that this procedure will enable a determination of the *actual* freezing point to be made, notwithstanding the fact that the water contains small

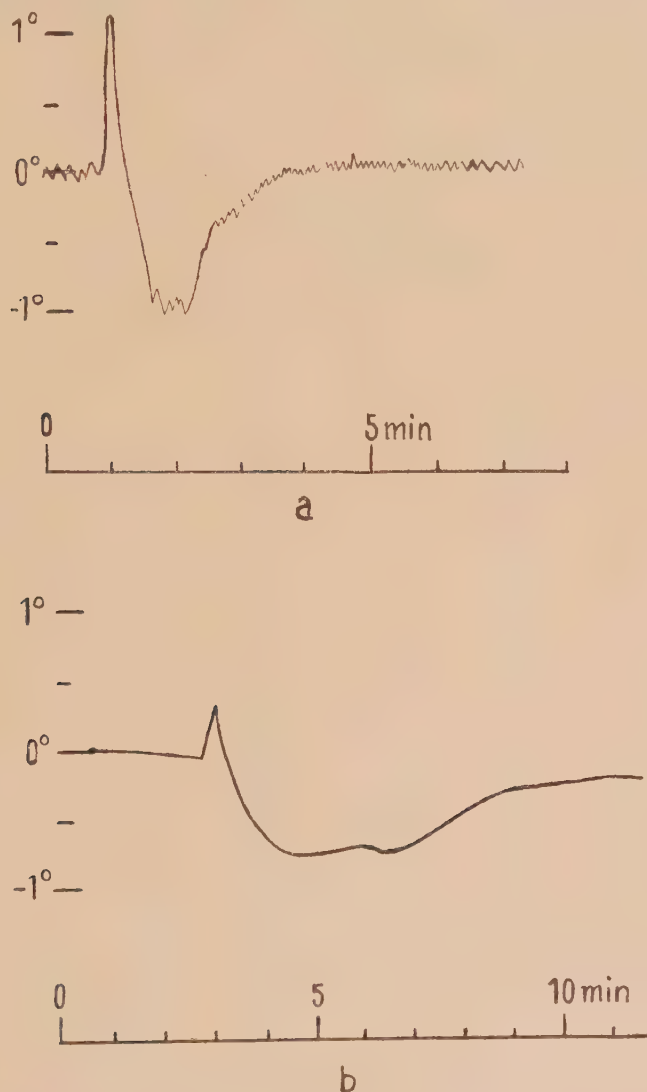


FIG. 5. Water surface temperature. Photographic registering curves. a. *Tömte*, 15. I. 1942, 21h. b. *Tömte* 8, 1. 1942, 12h.

quantities of solutable salts, which would lower the freezing point somewhat below 0° C. The determination of the actual freezing point is, however, of still greater importance in observations of the supercooling of running water, which require a higher accuracy of observation.

In order to start the freezing process afresh before observation, the ice-cover formed earlier was sometimes removed from the entire basin, and then the surface temperature registered by the radiation receiver, connected through a cable with the compensator and the photographic registering apparatus, which were placed in a small house quite near. At other times a plate of ice was cut out of the ice cover and submerged to a depth of about an inch, a small shallow basin with a bottom temperature of 0° C, and with a very slow flow of water, thus being formed. These conditions were favourable to the attainment of great supercooling. To study rapid freezing, some water was simply poured over the ice cover.

In the course of the observation period, 59 registering curves showed a temperature fall of the radiating surface below 0° C, during that phase of the freezing process when some open water was still to be found within the area limited by the effective space angle of the receiving cone. The shape of the curves was generally similar to the typical curves of Fig. 5 reproduced below, which were registered in January 1942 at a temperature considerably lower than that at Sörnesset. A summarized view of the observations is given in Table 1.

Table 1. *Observations of water surface temperature at Sörnesset 1941.*

n number of registrations. p limit numbers of supercooling period in minutes. t » » » minima of surface temperature (° C). T » » » air temperature (° C). v » » » wind (scale 0—12). N » » » cloudiness (scale 0—10). The observations reported are made before 7 ^h 30 M. S. T.						
	n	p	t ° C	T ° C	v	N
19. April 1941	8	1—5	—0.10 to —0.40	—5.6 to —4.5	0	0
23. April 1941	10	2—8	—0.12 to —0.66	—9.9 to —9.0	0	0
24. April 1941	2	1	—0.25 to —0.18	—7.6 to —7.7	0	0
25. April 1941	10	2—18	—0.22 to —0.39	—6.8 to —6.0	0	0
3. May 1941	5	2—6	—0.08 to —0.25	—5.7 to —4.2	0	0—1 CuStr
8. May 1941	3	6 >15	—0.11 to —0.70	—9.4 to —8.0	0	1
—»— evening	1	>10	—0.11	—2	0—1	8 ACu
9. May 1942	2	1 3	—0.12 —0.12	—8.0 —7.8	0	5 Cu
10. May 1941	6	2 >10	—0.27 to —0.55	—9.2 to —6.0	0	CiStr veil
12. May 1941	7	0.1—1.5	—0.03 to —0.13	—2.9 to —1.5	0	6—7 CiStr
14. May 1941	5	3—6	—0.24 to —0.49	—4.6	0—1	0—2

The period of supercooling varied between 0.1 and more than 18 minutes, depending upon the different factors influencing the amount of supercooling, which will be discussed below. The maximum of supercooling observed,—0.70° C, occurred on the 8th of May.

The experiences from these observations were used to improve the registering apparatus and observations were continued during December 1941 — January 1942 at *Tömté, Nannestad*, 275 m above sea level. Two wooden cylindrical vessels were placed in the open air, mounted on a platform between two horizontal wooden bars carrying the receiver (1), which might easily be removed from one vessel to the other. Generally, the observation started by placing the receiver over the first vessel containing water with an ice cover, the surface temperature of which will be below zero, proportional to the thickness of the ice, as seen below. The registering curve may thus start somewhat below the zero line. Then water of a temperature 1-2° C was poured into the second vessels and the receiver swiftly placed above, while registering was proceeding. The galvanometer registering curve (see Fig. 5) shows correspondingly, first a rapid rise and then from the sharp apex an exponentially falling curve corresponding to the cooling of the water surface. The curve passes below the 0° line to a broad minimum and then slowly rises again. During this part of the curve ice formation is started and is going on, till the curve approaches the 0° line again, a continuous thin ice cover then being formed. As representative types are reproduced two curves in Fig. 5. During the observation period 46 registering curves were obtained showing a temperature fall of the radiating surface below 0° C, the typical shape being that of the curves reproduced, although the period of supercooling may be much longer. A list of observations of the maximum of supercooling registered is given for a series of days in the table 2 below.

Table 2. *Observations of water surface temperature at Tömté 1941-42.*

	n	p	t ° C	T ° C	v	N
27. Dec. 1941	5	1.5	—0.66	— 17	0—1	0
18h—23h		to	to	to		
		3	—1.20	— 20		
28. Dec.	8	2	—0.60	— 21	0	10 Str
11h—20h		to	to	to		
		8	—1.0	— 17		
2. Janr. 1942	3	5	—0.58	— 11	0	0
20h—24h		to	to	to		
		7	—0.83	— 12		
6. Janr. 1942	5	2	—0.28	— 6	0—1	0
18h—24h		to	to	to		
		9	—1.2	— 13		
7. Janr. 1942	8	2	—0.39	— 15	0—1	1—10
9h—23h		to	to	to		StrCu
		>23	—0.90	— 7		
8. Janr. 1942	6	0.5	—0.10	— 12	0—1	1—10 Str
11h—22h		to	to	to		
		7.5	—0.78	— 9		
15. Janr. 1942	11	1.5	—0.25	— 20	0—1	0—10 Str
9h—22h		to	to	to		
		4	—1.06	— 16		

As pointed out before, the ice will grow from distinct starting points in the form of crystal beams, leaves, feathers, stars and other forms, being stretched out gradually. This process will leave open parts of the water between the growing elements of ice structure, till all the open parts are covered by an ice membrane, as thin as a leaf, but still showing some of the structural elements from which it has

(1) If the receiver is placed right above the second vessel while water is rapidly being poured into it, the vertical air movement caused will influence the thermopile, so this will have to be avoided.

been formed. During this period the average surface temperature of the area limited by the effective spacing angle of the receiver cone will be represented by that part of the curve (Fig. 5) situated below the zero line. Judging from the fraction of open water area in proportion to the ice area at proper intervals during each observation, it was found that *the negative temperature will be approximately proportional to the open fraction*. In order to see the consequence of this, we should remember that the surface temperature of the ice elements first formed will be very near 0°C , the thickness being in fact only a small fraction of a millimeter. This

is seen by considering the heat flow through an ice plate $S = -\lambda \frac{t}{H}$, t being the

temperature of the surface exposed to the loss of heat S , H being the thickness of the plate and λ the conductivity of ice $0.0051.3600\text{ cal/cm, hour, centigrade}$. In the paper previously mentioned formulae were given allowing a calculation of the heat loss S , when the meteorological conditions are known. At a moderate degree of cold, the value of S will be about $10\text{--}30\text{ cal-cm}^2\text{ hour}$. Taking for instance $S = 20$, and an ice thickness $H = 0.01\text{ cm}$, the surface temperature will be $t = -HS/\lambda = -0.01^{\circ}\text{C}$, which is within the limit of accuracy of observation (1).

Evidently, *the supercooling of the open areas of water remains unaltered, even while they are limited by ice* that is only gradually diminishing the open areas. In some cases this could be directly observed by moving the receiver slowly across neighbouring areas that are either open or covered by ice. The size and duration of the open areas, and consequently the amount of supercooling, will depend upon the number of nuclei present when the freezing process begins. If we deliberately suspend a considerable number, the open areas will be small and numerous, the supercooling will only reach a small value, and a continuous cover will be formed in a short time. If we remove the ice and clean the surface, the number of nuclei will be reduced, the open areas will be greater and will exist for a longer time. If the water keeps calm the supercooling will reach a greater value, and it will take longer time to produce a continuous ice cover. In return, however, the elements of ice structure first formed, will have grown thicker and will be seen as ripples on the ice cover.

An interesting example of this was presented by one of the registrations in Table 2, from Jan. 8th, 13h2^m—35^m. The minimum of temperature, -0.9° , was reached after 15 minutes, but a temperature of -0.4° was still registered after 22 minutes, when more than half of the effective area was covered by ice, chiefly by rather strong ice beams. In the small area still open some fine icewater was then added, which immediately caused the water to be covered by thin ice, while the registering curve simultaneously rose to 0°C . The temperature registered being the average value of the effective area, the actual supercooling of the small open area must have been several times greater than that registered (-0.4°).

Otherwise it should be emphasized that the actual supercooling will also depend upon the tranquillity of the air. A breath of wind which sets the water surface layer in motion will rapidly diminish the temperature gradient of the surface layer.

The influence of chance numbers of nuclei and accidental air conditions may thus cause considerable variations of the actual supercooling. This may sometime completely outweigh the influence of the general loss of heat, determined chiefly by air temperature and cloudiness. The actual observations of supercooling referred to in the above tables illustrate this.

It is of interest to note that the maximum of supercooling observed at Tömté was -1.2°C .

5. The observation series from the two winter periods treated in the preceding section, have shown that *the formation of ice in open water which is at rest or moving slowly, will regularly be preceded by a supercooling of the water surface layer*. The state of supercooling is stable in so far as supercooling may remain unaltered at a sufficient distance from solid phase boundaries. Such boundaries are, for instance the surfaces of nuclei suspended in the water from which the crystal growth starts. The transmission of heat from the zone of crystallisation will partly pass through the ice to the adjacent air, and partly to the open water which is supercooled, as indicated by Fig. 6. The transmission of heat however, is a process which necessarily needs time, depending upon the existing temperature gradients. The reason why

(1) Measurement of the surface temperature of an ice cover with subsequent measurement of the thickness, gives a simple method for determining the total heat loss from the ice surface.

the supercooling of water is generally regarded as a state which is unstable in itself, is probably due to the ordinary performance of demonstration experiments with marked supercooling, the crystallisation process proceeding at a very fast rate, approximately instantaneously, when starting nuclei are provided for. If the water is cautiously cleaned from nuclei, it may in fact remain in a state of supercooling for an indefinite period, even if it be exposed to violent, turbulent motion, as was pointed out by *Altberg* (1), in giving a detailed report of the investigations hitherto published on the subject.

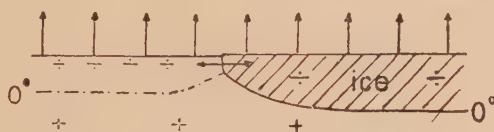


FIG. 6. Arrows indicate heat flow.

6. In section 1 it was pointed out that the formation of ice in running water *dynamic ice formation*, will be dependent upon the supercooling of the surface layer and its mixing with the deeper strata. If we could observe the actual field of movement, the field of heat current and the field of temperature in the water, we should find it very complicated when nuclei of crystallation are present. The growth of these will require a temperature gradient pointing outwards from the surface of each crystal, and the size of the gradient will again depend upon how rapidly the velocity changes with the distance from the surface. The growth will be greatest at those places where the field of velocity tends to increase the temperature gradient. This is the reason why sharp edges and points are not smoothed, on the contrary, the growth there will be great.

In rapidly running rivers, it is comparatively easy to observe a general supercooling of the water, provided the proper precaution are being taken. A favourable period for observation is either at the beginning or at the end of winter, when bottom ice is formed during the night. Solar radiation may then be sufficiently strong to loosen the bottom ice during the daytime, and a new bottom ice formation takes place the following night. At *Vosseelven* near *Skippe*, measurements were thus performed at Easter 1936 recording a temperature of -0.02° to -0.05° of the running water, while bottom ice was formed during a clear sky and atmospheric temperature -4° C. The measurements were performed with a mercury thermometer, division 0.01° .

In measuring the following precautions are to be taken :

- Thermometer division 0.01° , permitting readings to 0.001° .
- Thermometer to be placed in the shade also when in water.
- The mercury column will exert a pressure in the elastic glass bulb, depending upon the inclination of the thermometer; the same inclination to be used in all observations.
- To prevent ice formation on the bulb, the thermometer should be heated above 0° C before being placed in the water.
- Determination of the actual freezing point is made on the spot, the thermometer being in its fixed position, and a mixture of ice and water from the river itself being used.

In a subsequent study the formation of bottom ice as a practical problem at water power stations in winter will be treated.

For the researches here published the author is indebted to the *Chr. Michelsen Institute* and to *Kr. Birkelands Fond*.

(1) *W. Altberg*, Flusseis und Winterregime, V. Hydr.Konf. d. Balt.Staaten, Finland, Juni 1936, Bericht 7 D

KEEPING WATER FROM FREEZING BY MEANS OF COMPRESSED AIR

by

PENTTI KAITERA

In the winter of 1940 experiments were carried out in Finland with the object of keeping minor areas of water from being covered with ice. This was done by utilising the thermal reserve of water in such a way that compressed air was pumped into the water. When rising to the surface the air bubbles brought with them warm water from below and this water served to melt the ice cover. This method has been used in U.S.A. more than 30 years ago to protect the dams of water power plants from damages caused by the ice cover. In the following some results of experiments in connection with this matter are recorded together with some practical results achieved by the application of the method.

This method is entirely based on the utilisation of the thermal reserve of the water. By experiments carried out in laboratories and in the open air it has been established that water of even as low a temperature as $0.2-0.4^{\circ}\text{C}$ will melt the ice though the melting process is very slow. But when the temperature rises to 1.0° and more the melting effect increases rapidly and thus it is possible to find an economically profitable solution by which this method can be applied in practice.

During the winter 1940-1941 experiments were carried out in some lakes of the interior, partly also in the Gulf of Finland. The procedure was generally as follows: Compressed air stored in bottles was conducted from them through tubes to various depths under the ice cover. The air was ejected from the tube through a special nozzle and the speed of the outflow could be regulated by the size of the opening of the nozzle and the pressure in the tube. The melting effect was determined by establishing how many cubic meters of ice were melted within a certain time-unit — usually 24 hours. In the majority of experiments only one nozzle was used, but some of the experiments were carried out by simultaneous ejection through 2-4 nozzles. These were placed at a distance of about 4 meters from one another.

When the bottle of compressed air was placed on the surface of the ice the supply of air was conducted as follows: First a hole was drilled through the ice and through this hole the tube was conducted vertically down. The air then rose round about the hole through which it mainly escaped. When in some of the experiments air was conducted to such a point under the ice cover, where it did not escape, it formed under the ice air cavities which prevented the warm water rising from below from reaching the lower surface of the ice. Then the melting effect was considerably reduced. The melting effect of this procedure is thus depending on the fact that a continuous flow of warm water is allowed to touch the lower surface of the ice.

During the winter referred to 44 experiments of at least 24 hours duration were carried out. The results of these experiments are shown in Table 1. The depth of water at the place of observation varied in the various experiments from 2.5 to 18 metres. In some of the experiments the nozzle was placed on the bottom, in others some distance from the bottom. The table shows its depth during the various experiments. The volume of compressed air ranged between 1.4 and 12 litres per minute. During the experiments in the early winter the atmospheric temperature was exceedingly low, sometimes almost -40°C . During the end of the winter it was above freezing point. The temperature of the water during the various experiments varied considerably. As a rule the temperature of the water immediately below the ice was about 0°C . In the small lakes — such as Palojärvi and Tampaja — it was in the deeper strata above 3°C . Vanajavesi is a larger lake where the water is disturbed by the autumn gales at the time when it freezes. Here the temperature of the water was below 2.5°C during the time of the experiments. In the sea (Miessaari island) the temperature of the water was from immediately below the ice down to a depth of up to 5 metres below 0°C (-0.2°C) and only at the very bottom there was a small region where the temperature was $+0.4^{\circ}\text{C}$. Comparing the best methods for the determination of the influence of the temperature of the water on the effect of melting, it was found most suitable to use the temperature of the water at the depth of the nozzle as a standard. This temperature varied, as the table shows, from $0.4-4.2^{\circ}\text{C}$. The

Table No 1. Results of experiments during Winter 1940-41.

Number of experiment	Experimental site	Date when experiment	Depth of water on the observation point — in metres	Mouth-piece was sunk to a depth of — meter	Number of began mouth-piece	Volume of compressed air — litres/minute	Air temperature during the time of experiment	Water temperature at the site of mouth-piece	Thickness of ice in cm	Melted ice-volume during 24 hours—in cubic metres	
1	Palojärvi	29. 11. 1940	16	10	1	8.0	—15.0	4.0	4.0	31	16.0
2	»	6. 1. 1941	17	10	1	1.6	2.0	4.0	4.0	48	1.7
3	»	15. 1. 1941	13.5	10	1	4.4	—39.0	4.0	4.0	50	12.0
4	»	28. 1. 1941	17	5	1	2.0	—26.5	3.6	3.5	59	1.5
5	»	»	17	5	1	3.2	—26.5	3.6	3.6	58	1.9
6	»	»	17	5	1	4.1	—26.5	3.6	3.6	56	2.1
7	»	17. 1. 1941	17	10	1	1.4	—38.7	4.0	4.0	51	2.7
8	»	»	17	10	1	2.3	—38.7	4.0	4.0	51	4.4
9	»	»	17	10	1	2.8	—38.7	4.0	4.0	52	5.6
10	»	20. 1. 1941	17	10	1	3.6	—6.4	4.0	4.0	62	5.0
11	»	21. 1. 1941	17	10	1	3.6	—28.6	4.0	4.0	55	4.2
12	»	31. 1. 1941	17	2.5	1	2.3	—11.4	2.5	2.7	60	1.3
13	»	»	17	2.5	1	3.5	—11.4	2.5	2.8	58	2.0
14	»	»	17	2.5	1	4.6	—11.4	2.4	2.7	56	2.1
15	»	5. 2. 1941	3	2.5	1	2.7	—23.8	4.0	4.0	53	0.55
16	»	»	2.7	2.5	1	4.0	—23.8	3.9	3.8	51	1.06
17	»	»	3	2.5	1	5.3	—23.8	4.0	3.8	52	1.5
18	»	7. 2. 1941	7	5	1	2.3	—17.4	3.6	3.0	52	1.0
19	»	»	7.5	5	1	3.4	—17.4	3.6	3.5	51	1.7
20	»	»	6.5	5	1	4.5	—17.4	3.6	3.7	54	1.9
21	»	14. 2. 1941	17	10	1	2.0	—12.0	4.0	4.0	63	4.0
22	»	18. 2. 1941	17	10	1	6.5	—12.9	4.0	4.2	62	7.7
23	»	3. 4. 1941	18	10	1	3.1	—9.4	4.1	4.2	61	3.3
24	Vanajavesi	6. 3. 1941	11	10	2	8.0	—5.0	1.9	1.4	59	3.1
25	»	»	11	10	3	8.0	—5.0	1.9	1.4	67	4.8
26	»	»	11.5	10	4	8.0	—5.0	1.9	1.4	47	5.5
27	»	10. 3. 1941	16.5	16	2	6.0	—2.5	2.5	7	56	5.7
28	»	»	16.5	16	2	3.0	—2.5	2.5	2.6	59	2.8
29	»	11. 3. 1941	11	10	2	8.0	—7.5	1.7	1.6	56	2.6
30	»	»	6	5	2	11.0	—7.5	1.7	1.6	64	2.9
31	»	29. 4. 1941	8	8	1	9.0	—3.3	1.6	1.7	65	3.0
32	»	»	8	8	2	9.0	3.3	1.6	1.7	70	3.5
33	»	»	8	8	3	9.0	3.3	1.5	1.6	65	3.4
34	Tampaja	15. 4. 1941	9.5	7.5	1	9.5	1.1	2.7	3.6	67	6.0
35	»	»	10	7.5	2	9.5	1.1	2.7	3.6	68	5.6
36	»	»	10	7.5	3	9.0	1.1	3.4	3.7	67	11.8
37	»	16. 4. 1941	6	5	1	11.0	0.7	2.4	3.2	75	3.5
38	»	»	6	5	2	11.0	0.7	2.4	3.2	75	5.2
39	»	»	6	5	3	11.0	0.7	2.4	3.2	70	5.5
40	»	18. 4. 1941	2.5	2.5	1	11.5	0.8	1.8	2.2	76	0.85
41	»	»	2.8	2.5	2	12.0	0.8	2.2	3.5	70	2.12
42	»	»	3	2.5	3	12.0	0.8	2.2	2.5	71	2.4
43	Miessaari	6. 4. 1941	8	8	1	9.5		0.4	0.4	70	0.25
44	»	»	8	8	1	9.5		0.4	0.4	65	0.25

Table No 3. — *Altered group average values based on the values recorded on table No 2 when depth site of mouth-piece is 8 m.*

Compressed air volume. l/min.	Water tempera- ture at site of mouth-piece ° C.	Melted ice- volume m ³ /24 hours
8	0.4	0.2
8	1.6	2.7
8	2.4	4.2
8	3.6	8.7
4	1.7	1.4
4	2.4	2.3
4	3.7	4.1
2	2.5	1.5
2	3.9	3.0

The results are shown in Table 3 and Fig. 2. The average group-values in Fig. 2 show fairly accurately that the melting effect increases as the quantity of compressed air and temperature of the water increase. The increase of the quantity of air causes a strong current of water which quickens the transfer of heat from the water to the ice. It is evident that by using larger quantities of air the addition of a certain quantity of air causes more and more reduced increases of quantities melted. This was established by the experiments conducted between 28.3 and 3.4 1946. On these occasions an air quantity of 100 l-min was used. The results of these experiments are shown by the dash line in Fig. 2.

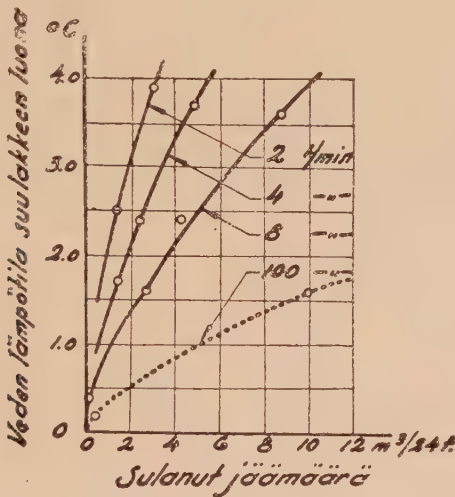


FIG. 2. Dependance of quantity of ice melted (m³/24 h) on quantity of compressed air and temperature of water at nozzle — depth 8 m.
Veden lämpötila = Temperature of water at nozzle.
Sulanut jäämäärä = Quantity of ice melted.

The figure also shows that by increasing the quantity of compressed air the same melting effect can be achieved up to a certain limit in cold as well as in warm water. For instance, by using a quantity of air of 8 l-min the same melting effect can be achieved in water of a temperature of 0.9° C as by using an air quantity of 4 l-min the temperature of the water being 1.5° C at the nozzle. The using of larger quantities of air is subject to greater initial- and running costs. Here the question arises

How to determine the limit for an economically profitable undertaking? The curves in Fig. 2 give an indication to the solution where this matter is concerned. When the temperature of the water is low the thermal reserve might prove inadequate to keep the water area from freezing during a cold period.

If there is available a compressor the effect of which is, for instance, $3.5 \text{ m}^3\text{-min}$ and which requires a motor of 35 H.P., a 20 m wide hole in the ice, the thickness of which is 0.5 m, can be worked under the average conditions stated in Fig. 2, on condition that the compressed air tubes are placed in the most advantageous places — approximately corresponding to the values shown in Table 4.

Table No 4. — *A compressor the effect of which is $3.5 \text{ m}^3/\text{min}$. and equipped with a 35 H.P. motor. To melt a 20 m wide crack in ice, thickness of ice being 0.5 m. Time for melting operation : 24 hours.*

Water temperature on a depth of 8 m. ° C.	Length of crack. m.	Theoretical motor effect to melt the crack. H.P.	Rate of machine effect (theoretical) compared with effect of 35 H.P. compressor
0.5	15	700	20
1.0	65	3100	90
2.0	160	7500	220
3.0	290	13500	390
4.0	450	21000	600

Table 4 also shows the amount of energy required for the melting of the openings in case all the energy necessary for the melting should have to be mechanically produced. Further, the summary accounts for the multiplied increase of the effect of the compressor in each case. When the temperature of the water is 0.5°C the gain of energy is 20-fold, at 2.0°C water it is 220-fold and at 4.0°C water 600-fold. This shows the advantage of this method with regard to economy of energy in comparison with methods in which the energy necessary for the melting is mechanically produced.

For the sake of finding out the influence of the depth of the bubbling point the observation material — not counting experiments in which the temperature of the water has been below 0.5°C or in which the number of nozzles has been 3 or 4 — has been grouped on the basis of depth on four groups and on the basis of the temperature of the water and quantity of compressed air in two groups when — by making further use of the curves in fig. 2 — the average group-values shown in table 5 are arrived at.

Table 5. Average group-values showing the influence the depth of nozzle, the temperature of the water being 4°C and 2°C and the quantity of compressed air 8 l-min and 4 l-min.

Water temperature at site of mouth-piece	Volume of compressed air. l/min.	Depth site of mouth-piece. m.	Melted ice- volume $\text{m}^3/24 \text{ hours}$
4	8	7.5	9.2
		10.0	13.2
4	4	2.5	1.1
		5.0	2.6
		10.0	7.2
2	8	2.5	1.0
		5.0	2.3
		9.0	3.7
		16.0	5.7
2	4	2.5	1.5
		16.0	2.8

The results are also shown in Fig. 3 where the points of the same quantity of compressed air and the same temperature of water are connected. The curves

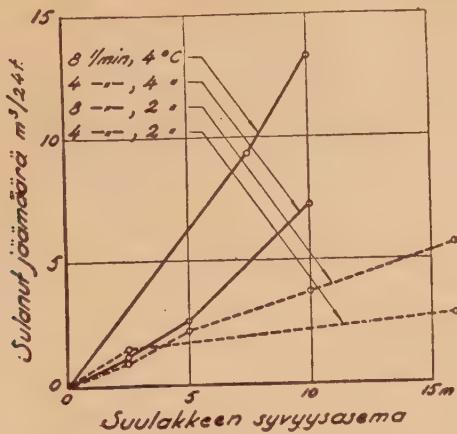


FIG. 3. Dependence of quantity of melted ice on depth of nozzle under various conditions.
Sulanut jäämäärä l³/24 t = Quantity of ice melted m³/24 h.
Suulakkeen syvyyssäema = Depth of nozzle.

show that when the bubbling point is lower the effect of melting grows greater the higher the temperature of the water is. When the quantity of compressed air is constant, the bubbles, when rising from a deeper position, move a greater mass of warm water than in shallow water, in which latter case the method thus offers less favourable prospects than in deep water. In deep water a greater amount of energy is certainly needed to eject the same quantity of air from the tubes than in shallow water but, on the other hand, a greater depth of water generally means a higher temperature and a greater thermal reserve in the mass of water which is set moving by the bubbles. It is also evident that in 10-20 m deep water the transfer of thermal reserve can take place from a further peripheral distance than in shallow water, which also increases the available thermal reserve.

The melting effect is also influenced by the number of nozzles used for the ejection of a certain quantity of air into the water. After certain adjustments the following data concerning the influence of the number of nozzles are arrived at :

Number of nozzles	1,	quantity of ice melted	3,5 m ³ -24 h
»	2	»	3,9 »
»	3	»	4,9 »
»	4	»	5.0 »

In each group the other average conditions are as follows :

Average depth of nozzles	6.6 m
Quantity of compressed air	9.7 l/m.
Temperature of water at nozzle	2.2° C

From this it can be noticed that the melting effect accomplished by the same quantity of air grows as the number of nozzles grows. The result tallies with the previously stated facts as larger masses of water are set in motion in this case.

Further, attention should be given to the influence by the temperature of the atmosphere. As shown in table 5, experiments have been carried out under as low a temperature as — 39° C. It is evident that such a severe frost will decrease the effect of the melting process as compared with a similar process during thaw. The observation material is, however, too inadequate to allow a numerical representation of the influence of temperature, as it depends inter alia on the thickness of the ice and the size and shape of the opening which varied considerably during the experiments. It is, all the same, to be noted that the temperature of the atmosphere is not a decisive factor as long as the surface to be melted is not too big. A

continuous keeping of a surface of water from freezing during a whole winter season makes the importance of the influence of the temperature grow. In the winter 1946-47, when an opening 30×80 m for floating timber was kept free from ice, the compressor had to be kept running 20-22 hours a day during hard frost but during light frost the time was reduced to 2-4 hours a day.

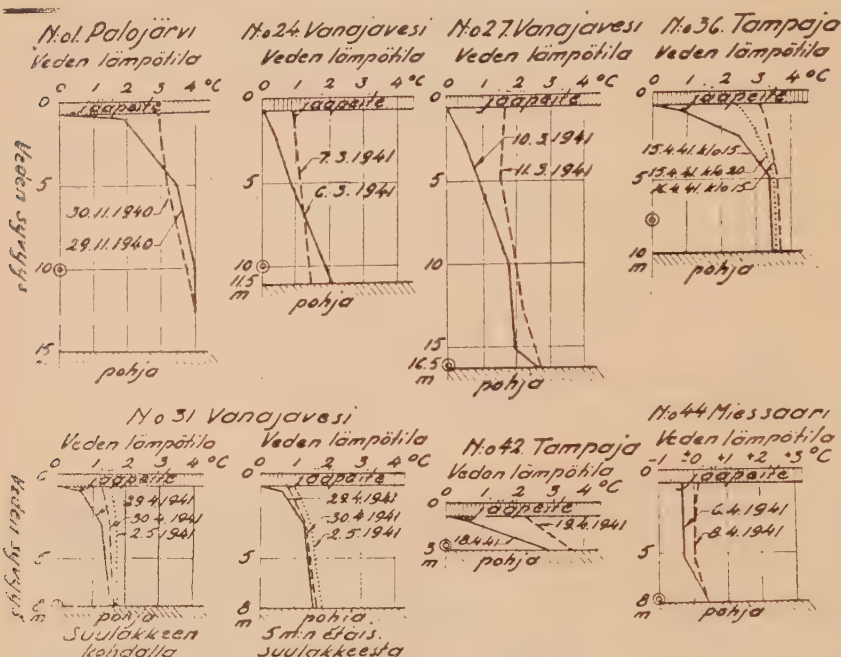


FIG. 4. Temperature of water at various depths before trial (drawn line), after 24 hours trial (dash line) and at various phases of the trial (dotted line). Double circle denotes depth of bubbling point.
 Veden lämpötila = Temperature of water.
 Jääpeite = Ice cover.
 Veden syvyys = Depth of water.
 Pohja = Bottom.
 Pohja suulakkeen kohdalla = Bottom at nozzle.
 Pohja 5 m:n etäis. suulakkeesta = Bottom 5 metres from nozzle.

During the experiments attention was paid to the fact that the temperature of the water near to the melting opening did not fall as much as could be expected considering the quantity of heat necessary for melting the ice and keeping the water from freezing. This did not happen even when the depth of water was remarkably shallow: 2.5-5.0 m. In order to demonstrate this fact Fig. 4 shows the temperatures of the water at the beginning and end of experiments with various depths (continuous lines). The depth position of the bubbling point is marked with a double circle. The experiments demonstrated in this figure are the same as those marked in Table 1 with corresponding numbers. As shown in the figure the temperature of the water of the strata, including at the depth of the bubbling point, has risen in many experiments independent of the fact that considerable quantities of heat has been required for melting the ice.

Table 1 shows that, taking into consideration the whole observation material, the temperature at the bubbling point has risen an average of 0.1°C . The number of experiments where the temperature of the water has remained unchanged or has risen has been 34, while a fall of temperature has taken place in 10 experiments. It is possible that in several cases there has been near the bubbling point hollows in the bottom from where warmer water has risen. The experiences which have been got later from practical applications of the method, confirm the idea that thermal reserves are transferred to such an extent from peripheral points that, under known conditions, there is no fear of their exhaustion.

Considering the conditions under which the method is applicable in practice, the temperature of the water during the winter months must be taken into account. In this respect the sea, the rivers and the lakes differ very much from one another.

During the early Winter the autumn gales mix the surface layers of sea water. The water remains a long time open and, as the temperature of it can fall below 0°C before any freezing takes place, because of the percentage of salt, the surface layers up to a depth of 30-40 metres is cooled to about the freezing point. Generally the temperature of this top layer is $0.1-0.3^{\circ}\text{C}$ and even below 0°C in the late winter. The wind also causes currents in the surface water. Cold water is forced into the bays of the shore and cools the water there down to the bottom also where the ice cover protects it from the influence of winds and cold air. Under these circumstances the conditions for keeping even small surfaces free from ice by the use of compressed air are not at hand. Exceptions do, however, exist. Thus temperatures of $1.0-2.0^{\circ}\text{C}$ at a depth of 5-10 m were recorded in some bays in connection with the experiments. It can also be assumed, that in such bays into which larger rivers discharge, the temperature of the water mainly depends on the thermal reserve of the water brought by the river.

In the winter of 1946-47 experiments were carried out to find if it was possible to transfere thermal reserves from deeper water through a colder surface layer. The experiments were arranged near the lighthouse of Söderskär. The temperature of the water down to a depth of 30 m was below 0°C . Lower down the temperature rose rather abruptly, being 1.3°C at 40 m and 3.2°C at 50 m. Through one nozzle 200 l/min. air was ejected. During the experiment currents appeared in the surface water in such a way, that the bubbles at times rose 10-20 m away from the opening above the nozzle. This as well as the depth of the cold surface layer caused the temperature of the water to rise only 0.1°C and could not bring about any melting of the ice.



FIG. 5.

The temperature of river water depends essentially on the situation of rapids. In these water comes into contact with the cold air and its temperature easily falls to about 0°C . In rivers, however, new thermal reserves are brought from the upper courses and there is no fear of exhaustion of the thermal reserve. By using sufficiently large quantities of air the method can be applied inter alia for the protection of the dams of power stations from damages caused by the ice.

In lakes the thermal reserves are essentially more favourable than in the sea or the rivers as far as the application of the method is concerned. Already at a depth of 5-10 metres the temperature of the water in the Finnish lakes usually is above 1.0°C , whereby the effect of melting grows high enough to make it possible to find paying solutions for practical purposes. Also in the lakes the gales of early

winter cause a mixing of the surface water. Because of this the temperature of big lakes is considerably lower than in small lakes. Investigations, have shown that in such lakes, where the area exposed to the wind is at least 3 km, the temperature of the water at a depth of 5 metres varied between 0.5-2.0° C during the winter months. In small lakes the measured variations at the same depth were 1.5-4.0° C.

Because the results of the experiments accounted for here above, the saw-mills on the banks of the Finnish lakes have started to keep the sorting basins open by using compressed air. The method has shown itself very economic and by applying it labour has been saved and the result of sawing has increased to such an extent that the cost of the plant has been paid within one winter. At the Ilola saw-mill on the shores of lake Päijänne measuring of temperatures was arranged in the winter of 1946-1947. to find out to what extent there was a danger of the thermal reserve being exhausted as a large surface had to be kept continuously melted. The size of the sorting basin of that saw-mill is 30 × 80 m (Fig 5) and the depth of water there varied between 4-8 metres. Air from the compressor is conducted through a net of tubes on the bottom. To keep the opening melted it has been necessary to use about 3.0 m³/min air. The performed measurements of temperature have shown that during the whole winter the temperature of the water in the melting opening as well as outside it remained constant at a depth of 3 metres i.e. 0.7-0.8° C. On the bottom the variations were somewhat greater, the temperature at the melting area being 0.2° C lower than at a distance to the side, where it remained 1.2° C the whole winter. The winter mentioned above was exceptionally cold and during January-March the temperature of the air varied between — 10 and — 30° C. During the days of hardest frost air had to be compressed up to 22 hours a day, but at a temperature of appr. — 10° C 10-15 hours work was sufficient. When the temperature rose to 1-4° of frost towards the end of March, there was no more need to use the compressor. These as well as other applications show that transfer of heat from peripheral areas takes place with the currents of water in such a way that there is no fear of the thermal reserve being exhausted.

5) **SUR LE PROBLEME DE LA REALISATION
DES PUBLICATIONS INTERNATIONALES CRYOLOGIQUES**
(monographie encyclopédique et revue)

par

le Prof. A. B. DOBROWOLSKI, Varsovie, Pologne.

La *nécessité* d'une encyclopédie cryologique et d'une revue cryologique internationale a été proclamée et motivée par l'auteur, depuis 1923 (1) dans plusieurs de ses publications, de même que devant des Congrès (2) et approuvée par la 4^e Conférence Hydrologique des Etats Baltiques, Leningrad, 1933 (3). Elle a été présentée comme une thèse à la dernière Assemblée Générale de l'Association d'Hydrologie Scientifique, Edimbourg, 1936, sa discussion ayant été reléguée à la 7^e Assemblée, Washington, 1939 (4).

Voici les *arguments* en faveur de cette thèse.

1. Il n'y a rien qui soit capable de lier entre eux les travailleurs d'un domaine de la Science autant qu'un *organe*. Les Congrès ne donnent qu'un contact éphémère et pas trop fréquent. Or plus le domaine est vaste, différencié, riche, et plus ses travailleurs ont besoin du contact le plus fréquent possible.

2. C'est l'organe qui contribue à assurer à l'ensemble des recherches la *continuité* de même qu'une *direction rationnelle* et un *plan*.

3. A part l'atmosphère, il n'y a peut-être pas d'autre grand domaine de la nature terrestre où l'*interdépendance* des différents phénomènes et différents facteurs soit aussi prononcée que dans la cryosphère, l'unité de la substance, plus homogène que celle d'un autre composant de la lithosphère, en étant la cause. Les différentes branches de la cryologie dépendent donc, dans leur travail, les unes des autres d'une façon *particulière*. Par conséquent, les différents spécialistes travaillant dans ces différentes branches ont *particulièrement* besoin de contact, de connaissance de ce qui se fait chez les voisins.

4. L'apparition de la glace dans tous les milieux : gazeux, liquides, solides ; l'influence que ce corps exerce sur les phénomènes météorologiques, géologiques, océanographiques, hydrographiques, biologiques ; enfin ses propriétés spécifiques, comme minéral et comme roche ; tout cela fait que ce corps a été étudié, depuis longtemps, par les spécialistes les plus divers ; nous y trouvons des mathématiciens et des astronomes, des physiciens et des chimistes, des cristallographes et des minéralogistes, des géographes et des géologues, des météorologistes et des climatologistes, des biologistes, de même que des ingénieurs et des touristes. Chacun publiait son travail ou ses notes dans une des revues consacrées à sa spécialité, dans des publications d'Académies, dans un récit ou un compte-rendu de voyage. Les travaux sur la glace se trouvent donc *dispersés* dans des publications de toutes sortes. Cela fait qu'il est difficile de s'orienter non seulement dans l'ensemble de ce domaine de recherches, mais même dans des conditions particulières. Très souvent on ignore exactement ce qui a déjà été fait et ce qui reste encore à faire ; on ignore aussi l'histoire du problème dont on s'occupe. Cet état de choses va en s'aggravant à mesure que l'étude de la glace se développe, que le nombre des travaux s'accroît.

(1) A. B. Dobrowolski, Histoire naturelle de la glace (Historia naturalna lodu), Varsovie 1923, Supplément français, paragraphe 2, pp. 922-923.

(2) Sur la nécessité de la fondation d'une institution internationale pour l'étude de la glace, 4^{me} Conférence Hydrologique des Etats Baltiques, rapport N. 97, Leningrad 1933, pp. 3-5. — Cryosphère, cryologie, cryotechnique et la question de l'organisation internationale des études de la glace et des phénomènes liés à la glace-Bull. de la Sc. Géophysique à Varsovie, fasc. 9-10, Varsovie 1934, p. 63. — Sur la problème de la réalisation d'une organisation cryologique internationale, 6^e Assemblée de l'Association d'Hydrologie Scientifique, U.G.G.I. Edimbourg 1936, Comptes Rendus et Mémoires des Commissions des Neiges et des Glaciers, Bull. 23, pp. 65-75.

(3) Comptes Rendus, Séance plénière du 14 sept., Leningrad 1933, p. 238.

(4) Bulletin 22, p. 636, point 2.

Deux choses seraient donc d'une grande utilité pour les différents spécialistes travaillant dans la cryologie.

Ce serait, d'abord, une *monographie encyclopédique*, aussi complète que possible, une vue d'ensemble donnée par un exposé systématique et détaillé de tous les problèmes relatifs à la glace, dans leur évolution et leur état actuel, avec une bibliographie et un matériel iconographique suffisants. Ce serait, ensuite, un *organe international*, qui donnerait aux investigateurs la possibilité de lire les travaux des autres.

2

Pour le *plan* de ces deux publications, je me permets de proposer ce qui suit.

1. A. Pour la monographie encyclopédique

La *monographie encyclopédique* pourrait bien prendre, comme base, mon livre « Histoire naturelle de la glace » (traduite dans une langue internationale), lequel présente justement le premier essai de synthèse des recherches faites dans la nature et dans le laboratoire, sur la glace de tout aspect et de toute origine, en même temps qu'une sorte d'index de tous les problèmes relatifs à ce corps si important et si peu connu encore, avec un exposé de l'histoire de chacun de ces problèmes, des résultats acquis, des questions litigieuses et des lacunes à remplir — comme le montre la table des matières très détaillée, en français, jointe à la fin du Supplément français, et en anglais, à la fin de la traduction anglaise des chapitres XII et XIV (« Théories de la structure et du mouvement des glaciers ») de mon livre précité.

Pour être une monographie encyclopédique désirée, mon livre offre les défauts suivants :

a) Il est publié en 1923. En particulier, la bibliographie (2000 titres environ) n'est pas complète pour la période de 1916 à 1923. Le livre est donc *suranné*, bien qu'il soit toujours actuel dans la plupart de son texte (1), et qu'il présente l'avantage de tracer l'histoire des problèmes cryologiques jusqu'à la période précitée.

b) Une monographie encyclopédique complète à tous les points de vue dépassant les forces d'un homme, j'ai été obligé d'en réduire le volume. Je n'ai pas traité toutes les questions avec la même importance. Je n'ai point insisté sur celles qui ne dépendent que d'une façon indirecte des propriétés spécifiques de la glace, ou bien qui font l'objet de monographies excellentes, détaillées et bien connues. En outre, le livre traite de la *substance* du H_2O solide, plutôt que des phénomènes qui y sont liés et qui en dépendent (p.ex. de la glaciologie il ne parle pas de la classification, de la structure et du mouvement des glaciers). Il ne parle rien que de la *Cryotechnique* : de l'utilisation de H_2O solide et de la lutte contre cet élément.

c) En traitant mon livre non seulement comme une encyclopédie raisonnée, mais aussi comme une sorte de *manuel*, j'ai classé le matériel en chapitres en me mettant non seulement au point de vue d'une classification scientifique, mais aussi à celui de la *facilité de l'exposé*. Ce qui a fait qu'au lieu de commencer par les propriétés générales de la glace et d'en faire les premiers chapitres, j'ai partagé l'étude de toutes ces différentes propriétés parmi les différents chapitres. De plus les questions cryologiques sont traitées sur le fond des problèmes plus généraux dont elles ne constituent que des cas particuliers, ce qui a donné certes, à ces questions la place et l'importance qui leur conviennent, de même qu'une netteté et compréhensibilité, mais, en revanche, a augmenté fortement le volume du livre, et serait, évidemment superflu dans une monographie encyclopédique destinée aux cryologues.

d) Le plus grand défaut du livre, c'est d'être *publié dans une langue qui n'est pas internationale* (2).

Malgré tous ces défauts, l'Histoire naturelle de la Glace « pourrait, peut-être, avec des monographies partielles d'autres auteurs (Hess, glaciologie; Seligman, la neige, particulièrement celle des montagnes; H.P. Barnes, « Ice Engineering »; Weinberg, rédacteur, livre « Glace » (Léd.), en russe, 1940, trad. Partielle de l'« Histoire de la glace », complétée des nouveautés d'applications à la pratique) faciliter la composition de la monographie encyclopédique désirée.

(1) Ce qui montre l'extrême lenteur du progrès de la cryologie !

(2) C'est que les fondateurs de la « Caisse Mianewski », institut polonais pour l'avancement des sciences, fondé avant la première guerre mondiale et qui était mon éditeur — de peur d'une intervention russifiatrice du gouvernement russe d'alors, avaient fini par introduire dans leur statut une clause défendant de publier en langues étrangères. Au moment de la publication de mon livre, la clause n'a pas encore été abolie.

2. A mon avis, bien que je voudrais bien voir la langue française comme la seule langue internationale, la monographie encyclopédique en question devrait être publiée en anglais, vu que tous les cryologues actuels savent sans doute lire dans cette langue.

3. Une monographie encyclopédique *complète* ne saurait être l'oeuvre d'un seul homme, elle sera donc une oeuvre collective. Pour le choix des rédacteurs, il faudrait procéder à l'élection d'une commission qui s'en occuperait. Aux rédacteurs choisis on laisserait le choix des collaborateurs. On trouvera, certainement, pour chaque problème, des spécialistes suffisamment compétents.

B. Pour une revue internationale (en langues internationales)

La revue cryologique internationale, je l'imagine comme un *mensuel* divisé, en *Mémoires* et *Bulletin* lequel comprendrait : articles sur les problèmes courants discussions, procès-verbaux des séances des associations cryologiques nationales, bibliographie, revues des publications parues, varia.

3

Ce n'est pas ici la place pour calculer les *frais approximatifs* de publication. Nous n'avons qu'à donner une base, assez facile à prévoir, pour un tel calcul.

Pour la *monographie encyclopédique* — à rééditer à peu près tous les six ans — on aura, pour la première édition : texte, bibliographie (environ 4000 titres), illustrations et graphiques, en tout 3000 *pages grand in-8°*.

Pour le mensuel, un mémoire par mois (pour commencement) de 32 pages en moyenne, bulletin 80 pages, en tout 7 *feuilles grand in-8° par mois*.

Ce que je crois certain — et je ne cesse pas de la répéter — c'est que restant *Commission* nous n'aurons jamais d'argent pour ces dépenses ; remarquons que nous devons dépenser *en plus*, tous les trois ans, une somme assez considérable — qui n'ira probablement pas en diminuant — pour l'impression des rapports présentés à la Commission pour les Assemblées générales de l'Association.

Peut-être réussirait-on à recevoir, en attendant, des fonds nécessaires, en s'adressant à L'UNESCO.

REMARQUE

Le présent rapport a été envoyé à la 7^e Assemblée de l'Association d'Hydrologie Scientifique, Commissions des Neiges et des Glaciers, Washington 1939.

par

le Prof. ANTOINE BOLESŁAS DOBROWOLSKI, Vasrovie, Pologne.

La VII^e Conférence Hydrologique Baltique, tenue à Lubeck et à Berlin en 1938, a accepté ma thèse affirmant l'utilité d'une *collaboration constante des hydrologues avec les météorologues, des services hydrologiques avec les météorologiques*.

Vu son caractère général, je me permets de la soumettre aux délibérations et aux votes de l'Assemblée de l'Association d'Hydrologie Scientifique, la thèse ayant déjà été présentée au Congrès de Washington (voir News Paper, organe de la Commission des Neiges et des Glaces, June 1, 1939, Post-scriptum, pp. 1-3), sans être pourtant mise aux délibérations, ce qui m'oblige de la répéter.

La thèse consitue une généralisation d'une autre thèse votée, elle aussi, par la Conférence de Lubeck et qui en présente un cas particulier, notamment le cas de l'étude des crues.

Voici, en résumé, les raisons de cette généralisation — raisons banales, non seulement parce qu'elles sont bien connues de tout hydrologue et de tout météorologue, mais encore parce qu'elles se résument toutes en une seule, notamment que *le domaine d'hydrologie est intimement lié avec celui de météorologie — plus intimement qu'avec tout autre domaine de recherches* :

1. Dans leur travail, les deux services, qu'ils soient officiellement séparés (le cas le plus fréquent) ou unis (Suède, U.R.S.S.), qu'ils ressortent à un même Ministère ou à deux Ministères différents, sont toujours intimement liés et dépendants l'un de l'autre.

En effet, sans connaissance des données météorologiques dans le bassin donné : précipitations, donc humidité et température, donc vents et nébulosité, donc distribution de la pression atmosphérique dans l'espace et dans le temps, l'hydrologue ne saura utiliser convenablement ni observations des hauteurs limnimétriques, ni mesures du débit ; l'ingénieur hydrotechnicien aura des difficultés sérieuses dans son travail ; on ne saura calculer le cadastre des forces hydrauliques, ni s'orienter dans l'économie aquatique du territoire donné, ni calculer le coefficient d'écoulement, ni prévoir le moment de congélation des rivières.

Il est vrai que si la météorologie est indispensable pour l'hydrologie, la relation inverse n'est pas si intime. Cependant, sans connaissance des données hydrographiques, un climatologue n'est pas à même d'élaborer convenablement ses synthèses prescrites, car il lui manque certains éléments très précieux, bien que de caractère plutôt auxiliaire.

2. Il y a toute une longue série de processus et d'événements naturels très généraux et très importants, lesquels touchant l'eau en même temps que l'air, intéressent également l'hydrologue et le météorologue — tels que l'évaporation des eaux, et le processus inverse ; le moment et la façon de la congélation des eaux, de la débâcle des glaces, de la fonte des neiges, etc. ; ici, les deux domaines ne sont plus à distinguer et les travaux respectifs sont tantôt faits par les hydrologues, tantôt par les météorologues.

3. Aux services météorologiques ressort la précision de certains événements hydrographiques. D'un autre côté, on voit souvent un service hydrographique fonder son propre réseau de stations météorologiques, de même que des services météorologiques organiser à leur soin une section hydrologique. Ainsi par exemple, la *Bayerische Landesanstalt für Gewässerkunde* « organisa, pour ses besoins, une section spéciale de météorologie hydrologique » (*Hydrometeorologische Abteilung*) et engagea pour son chef, M. Hauser, météorologiste bien connu ; grâce à quoi, on a pu constater une amélioration notable des pronostics relatives à la hauteur des cours d'eau. Enfin il n'est point rare de trouver un hydrologue et un météorologue en la même personne (Finn Malmgren, Jean Lugeon).

Telles sont, en résumé incomplet, les raisons pouvant justifier ma thèse — raisons, peut-être, superflues, la thèse paraissant trop évidente.

A la même VII^e Conférence Hydrologique Baltique j'étais allé plus loin, en affirmant que les meilleures conditions pour la *collaboration constante* des deux services seraient données par leur union officielle en un *service* hydro-météorologique. Je ne veux pas répéter ici les motifs, faciles à trouver. Je mentionnerai seulement une

preuve de fait : c'est que cette union a été réalisée, après la guerre mondiale, en Suède et dans les Soviets — deux pays où l'hydrologie de même que la météorologie avaient atteint avant la guerre un haut degré d'organisation et où cette nouvelle expérience n'a pas donné de mauvais résultats.

Naturellement, on ne saurait réaliser partout et de suite ce postulat sans s'exposer à des dangers, vu les conditions et les circonstances particulières dont il faut tenir compte dans chaque pays, et c'est pourquoi cette seconde thèse n'a pas été acceptée. La condition sine qua non du succès d'une telle union est : que *les deux services soient suffisamment mûrs*, qu'ils atteignent un niveau matériel, scientifique et pratique suffisant ; dans le cas des niveaux trop différents, ou bien quand les deux services ne sont pas suffisamment mûrs, surtout quand ils ne sont qu'au début de leur développement, l'union ne sera que préjudiciable. Mais on devrait *tendre* vers une telle union pour la réaliser à un moment, différent pour chaque pays, où cette réalisation pourra se faire sans avoir rien à craindre, et c'est pourquoi je me permets de soumettre aussi cette seconde thèse aux délibérations et aux votes de l'Assemblée de notre Association.

TABLE DES MATIÈRES

DU TOME II

Commission des Neiges et Glaciers

Comptes-rendus des séances	5
I. Adresse du Dr Church, Président	13
II. Rapports relatifs aux Icebergs	17
1) J.J. SCHELL.—On the distribution of Icebergs in the Northern Hemisphere with special reference to South of Newfoundland	19
2) J. DE SAMPAIO FERRAZ. — Northern Sea Ice and Rainfall in S.E. Brazil	22
III. Caractéristiques physiques de la Couverture de Neige.	
Rapport Général de V.V. KORHONEN	25
Rapports nationaux :	
1) V.V. KORHONEN (Finlande). — Report of the Researches concerning Snow- and Iceconditions which have been made in Finland after 1936	31
2) R. JURVA (Finlande). — Sur les conditions de glace de la Baltique le long des côtes de Finlande	33
3) PENTTI KAITERA (Finlande). — Snow melting in the Spring and its effects on the opening of the Waterways in Finland	38
4) V.V. KORHONEN (Finlande). — Correction des mesurages des chutes de neige faite sur la base de la nature de la place de mesurage	39
5) W.N. Mc CLEAN (Great Britain). — Flooding due to the melting of Snow and Ice	41
6) R.W. GERDEL (U.S.A.). — Physical changes in Snow-Cover, leading to Runoff, especially to Floods	42
7) M. DE QUERVAIN (Suisse). — Ueber den Abbau der Alpinen Schneedecke	55
8) F.L. RHODES et W.T. WILSON (U.S.A.). — The cooperative Snow Investigation Program. Its objectives and operations	69
9) O.H. HOOVER (Canada). — Effects of Chinook (Foehn) Winds on Snow Cover and Runoff	86
10) G. HEINSHEIMER (Argentine). — On the relation of precipitation accumulation and melting of snow to the Stream flow in the San Juan River	101
11) Observations of Himalayan Snowfall and Snow-Cover made by the India Meteorological Department	109
12) J.E. CHURCH (U.S.A.). — The Evolution of Snow-Melt by Dyes and Drip-pan	115
13) J.E. CHURCH (U.S.A.). — Standardizing Method of Snow Cover Measurement	118
14) R.F. LEGGET. — Canadian Interest in Snow and Ice Research	128
15) G.J. KLEIN. — Canadian Survey of physical characteristics of snow covers	138
16) R. SMITH JOHANSEN (U.S.A.). — Snow Data	153
17) V.J. SCHAEFER (U.S.A.). — The development of a Snow classification system	157

18) W. MÖRIKOFER (Suisse). — The dependance on altitude of the snow cover in the Alps	161
19) E. BUCHER (Suisse). — Sur l'activité de l'Institut fédéral pour l'étude de la neige et des avalanches au Weissfluhjoch sur Davos	171
20) M. PARDE (France). — Influence de la couverture nivale sur le ruissellement et spécialement sur les inondations	175
21) R.E. LUNDQUIST. (U.S.A.). — Descriptive and quantitative classifications of various conditions persisting in snow and ice deposits	182
22) V.J. SCHAEFER (U.S.A.). — The formation of ice crystals in the laboratory and in the atmosphere	186

IV. *Rapports relatifs au grain de glacier*

1) A. RENAUD (Suisse). — Contribution à l'étude du grain de glacier	213
2) G. SELIGMAN (Great Britain). — The growth of the glacier crystal	216
3) H.W. AHLMANN and E.G. DROESSLER. — Glacier Ice crystal measurements at Kebnekazse. — Sweden	221

V. *Glacier Measurements*

J.M. WORDIE (Great Britain). — General Report	231
Rapports nationaux	
1) G.L. MERCANTON. — Rapport sur les variations des glaciers de 1935 à 1946 (Suisse, Italie, France, Autriche)	233
2) Variation des glaciers en Suède, Islande et Norvège	257
3) H.W. AHLMANN (Suède). — Recent glaciological Investigations in Sweden	262
4) V. MEEK (Canada). — Glacier Observations in the Canadian Cordillera	264
5) N.E. ODELL. — Glacial Features of the Lloyd George Mountains (British Columbia)	276
6) FIELD (U.S.A.). — The variations of Alaskan Glaciers (1935-1947)	277
7) H.J. SPANN. — Report on Glacier Recession in Peru	283
8) OPPENHEIM. — Deglaciation in Columbia	285

VI. *Recherches sur les glaciers*

1) G. SELIGMAN. — Report on the Progress of Glaciological Research in Great Britain	291
2) K. FAEGRI (Norvège). — On the variations of Western Norwegian glaciers during the last 200 years	293
3) GORDON MAULEY (Great Britain). — Glacier Variations and climatic fluctuations in Britain	304
4) R. HAEFELI and P. KASSER (Suisse). — Observations in the Firn and Ablation Regions of the Great Aletsch Glacier	306
5) R. BILLWILLER. — Die Kurzfristigen Schwankungen der Alpengletscher und ihre Ursachen (Suisse)	323
6) G.J. HEINSHEIMER (Argentine). — Horizontal Age Differences in Glaciers	329
7) E.A. MARCHETTI. — Studies on Snow and Glaciers in the Argentine Republic	332
8) R. STREIFF-BECKER (Suisse). — Der Wassercabfluss in einem Gletschertal	338
9) J.S. FISHER (U.S.A.). — The pressure melting point of ice and the excavation of cirques and valley steps by glaciers	341
10) J.E. FISHER. — Id.	345

11) A. RENAUD et P.-L. MERCANTON (Suisse). — Les sondages sismiques de la commission helvétique des glaciers	346
12) W. AHLMANN. — Preliminary glaciological Plans for the Norwegian-British-Swedish Antarctic expedition 1949-52	353

VII. *Formation de la glace sur les lacs et les rivières*

1) O. DEVIK (Norvège). — Ice Formation in Lakes and Rivers	359
2) J. LAMBOR (Pologne). — La genèse de la glace flottante et son apparition sur les cours d'eau de l'Europe centrale appartenant au bassin baltique	367
3) O. DEVIK (Norvège). — Supercooling and Ice Formation in open waters	380
P. KAITERA (Finlande). — Keeping water from freezing by means of compressed air	390
A.B. DOBROWOLSKI (Pologne). — Sur le problème de la réalisation des publications internationales cryologiques	399
Id. — L'hydrologie et la météorologie	402

